Volume 4, Number 4 July-August1996

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

Official English Translation of *Stratigrafiya, Geologicheskaya Korrelyatsiya*

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Boris S. Sokolov

Academician, Russian Academy of Sciences
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Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

МАИК НАУКА/INTERPERIODICA PUBLISHING

Stratigraphy and Geological Correlation

(Stratigrafiya. Geologicheskaya Korrelyatsiya)

ISSN: 0869-5938

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P.O. BOX 1831, BIRMINGHAM, AL 35201-1831 USA
PHONE: (800) 633-4931; (205) 995-1567 (outside US and Canada); FAX: (205) 995-1588

PUBLISHER
МАИК НАУКА/INTERPERIODICA PUBLISHING
PROFSOYUZNAYA UL. 90, MOSCOW, 117864 RUSSIA
PHONE: 7-(095) 336-1600; FAX: 7-(095) 336-0666

SUBSCRIPTION PRICES (1996, Volume 4, 6 issues)
US $699.00 (North America), $805.00 (other). A 5% discount is given on 2-year orders. Subscriptions may be entered at any time. Orders may be placed at agencies or directly. 5% discount to all subscription agencies. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Checks payable to *Stratigraphy and Geological Correlation* or Interperiodica.

Stratigraphy and Geological Correlation (ISSN: 0869-5938) is published bimonthly for (US dollars) $699.00 (North America), outside North America $805.00 per year by MAIK Nauka/Interperiodica Publishing, 5724 Hwy. 280 East, Birmingham, Alabama 35242-5724 USA. Second Class Postage Paid at Birmingham, Alabama. POSTMASTER: Send address changes to *Stratigraphy and Geological Correlation*, P.O. Box 1831, Birmingham, Alabama 35201 USA.

Journal founded in 1993.

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Some of the Geochemical, Biochemical, and Biotic Consequences of Eustatic Oscillations

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Received February 2, 1994; in final form, August 16, 1994

Abstract—Regressions and transgressions were caused by short-period sea-level fluctuations that created conditions for more intensive transportation of large amounts of biophilic elements from land to the water basins. This process considerably increased plankton productivity, especially that of shell-less phyto- and bacterioplankton, some of the species of which, under certain conditions, emitted toxins that often stunted or killed the calcareous organisms. Consequently, the sediments accumulated proportionally larger amounts of organic matter with a declining carbonate content. The concentration of numerous toxic elements in sediments, enriched with organic matter, and the diagenetic H_2S generation, sometimes accompanied by the hydrosulfuric poisoning of bottom waters, had a negative effect on benthic fauna. Therefore, in water basins, some of the biotic crises could have been caused by the simultaneous occurrence of interrelated events responsible for the stress effect on plankton and benthic organisms.

Key words: *sea-level oscillations, regression, transgression, sapropelites, biotic events, hydrosulfuric poisoning.*

INTRODUCTION

Changes in Phanerozoic marine biota were often concurrent with eustatic fluctuations that differently affected the ecological niches of various biota representatives, the temperature and salinity of sea water, the oxygen content in it, the circulation system of currents, and other environmental parameters (Kauffman, 1986; *et al.*).

As we know, the state of biota is greatly influenced by anoxic environments during the accumulation of deposits rich in organic matter (OM). This effect becomes particularly apparent, if we correlate the time of anoxic conditions in water basins with the distribution of the number of species of certain organisms in the Cretaceous: the anoxic episodes coincided with a considerable reduction of foraminifer diversity, whereas during the anoxic decline, these organisms flourished and appreciably increased the number of their species (Fig. 1) (Gorbachik and Kopaevich, 1992). In these periods, the geochemical conditions of sedimentation became fundamentally different from those before and after the episodes. The Cenomanian and Turonian boundary layers in the Crimea provide data on the specific geochemical and paleontological features of sediments formed at that time (the Aksudere geological section near Bakhchisarai). These deposits show a sudden change in the content of planktonic and benthic foraminifers and in the concentration of many chemical elements (Fig. 2). A similar pattern is observed in the Cretaceous deposits at other sites (Arthur *et al.*, 1990; Brumsack, 1986; *et al.*). These kinds of episodes are also typical of certain Phanerozoic time intervals (Jenkyns, 1988; *et al.*).

To understand the nature of these sudden biotic and geochemical anomalies, we have to know the cause-and-effect relations between them and the eustatic oscillations in order to reveal the possible mechanism of sediment enrichment with organic matter, and to find correlations of this phenomenon with the changes in biota.

REGRESSIONS AND TRANSGRESSIONS AS THE POSSIBLE CAUSE OF THE FORMATION OF HORIZONS RICH IN OM

Brockamp (1942) and Strakhov (1947) remarked on the connection between eustatic oscillations and the appearance of certain geochemical anomalies in sedimentary layers, for example, in the form of ferruginous formations. Hallam and Bradshaw (1979), giving no details, arrived at the conclusion that the transgressions, usually associated with short-period eustatic fluctuations of the third order, are often accompanied by the accumulation of sediments rich in OM.

The results of the study of organic matter composition testify that, along with the marine mainly planktonogenic OM, the majority of the Mesozoic and Cenozoic black shales also contain allothigenic OM supplied from land (Simoneit, 1986; *et al.*). These circumstances should be taken into account when studying the possible causes of the black shale-type formation of carbonaceous deposits.

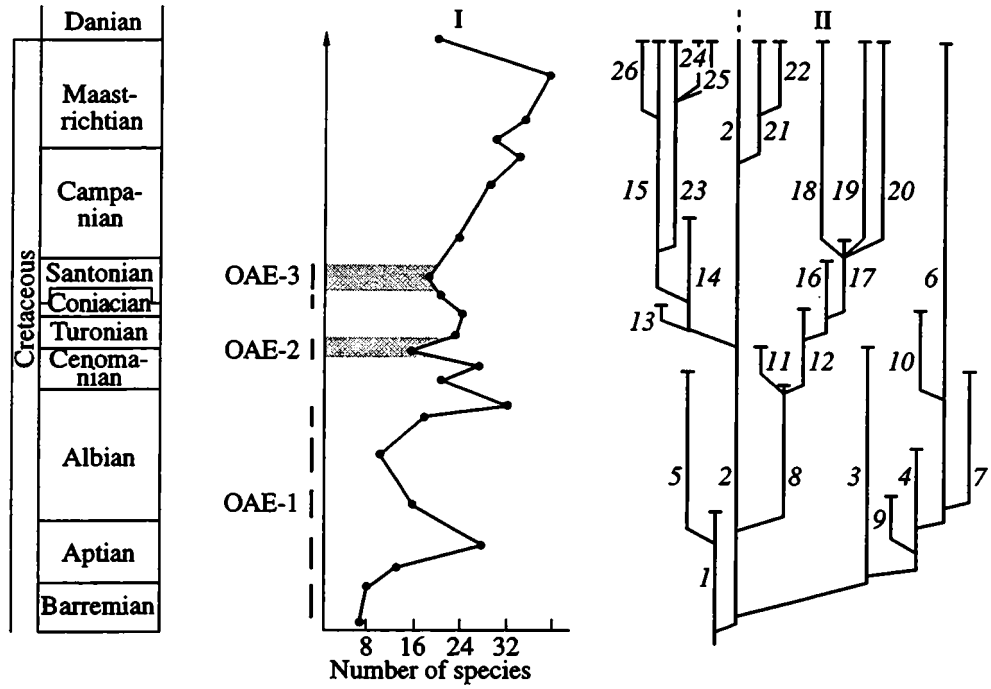

Fig. 1. Schemes of taxonomic diversity of species (I) and philogenetic evolution of planktonic foraminifer genera (II) in the Cretaceous.

Oceanic anoxic events: (OAE-1) Barremian—Albian; (OAE-2) Cenomanian—Turonian; (OAE-3) Coniacian–Santonian (?). Taxa: (*1*) *Globuligerina*; (*2*) *Hedbergella*; (*3*) *Clavihedbergella*; (*4*) *Blowiella*; (*5*) *Favusella*; (*6*) *Globigerinelloides*; (*7*) *Planomalina*; (*8*) *Ticinella*; (*9*) *Leopoldina*; (*10*) *Hastigerinoides*; (*11*) *Rotalipora*; (*12*) *Praeglobotruncana*; (*13*) *Helvetoglobotruncana*; (*14*) *Whiteinella*; (*15*) *Archaeoglobigerina*; (*16*) *Dicarinella*; (*17*) *Marginotruncana*; (*18*) *Globotruncanita*; (*19*) *Contusotruncana*; (*20*) *Globotruncana*; (*21*) *Globotruncanella*; (*22*) *Abathomphalus*; (*23*) *Rugoglobigerina*; (*24*) *Rugotruncana*; (*25*) *Kuglerina*; (*26*) *Bucherina*.

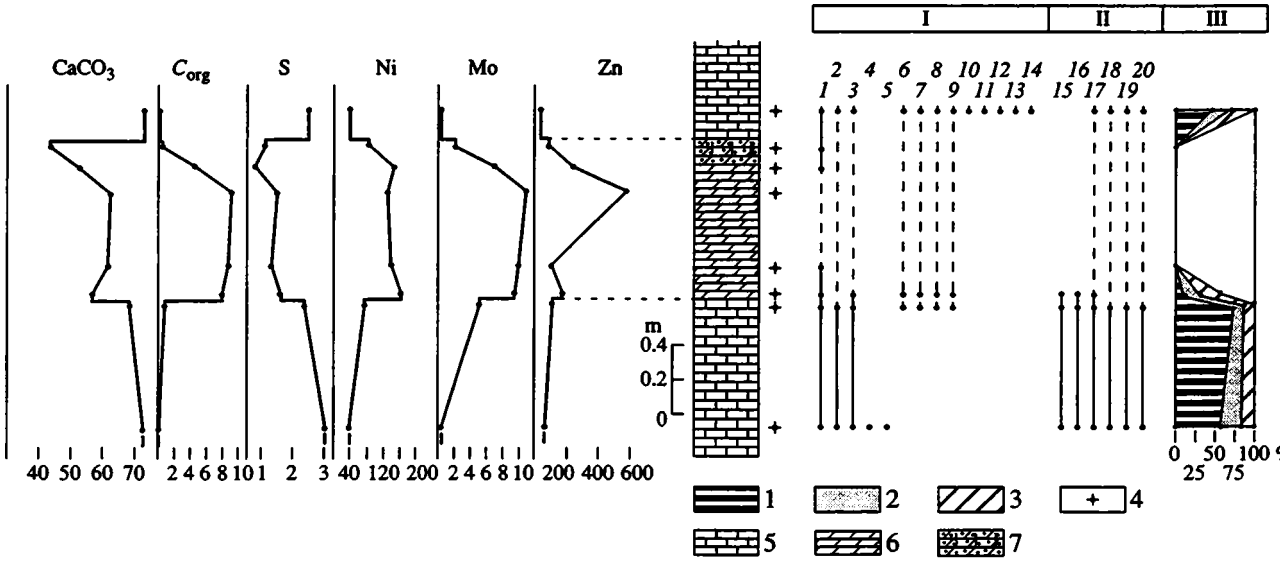

Fig. 2. Lithological column of Cenomanian—Turonian boundary layers (Crimea, Aksu-Dere locality) and distribution of chemical elements and various foraminifer species.

$CaCO_3$, C_{org}, and S contents are given in wt %; Ni, Mo, Zn in ppm (recalculated for carbonate-free matrix); I, planktonic foraminifers: (*1*) *Hedbergella* spp., (2) *Dicarinella* cf. *algeriana*, (*3*) *Whiteinella brittonensis*, (*4*) *Rotalipora greenhornensis*, (*5*) *R. cushmani/deeckei*, (*6*) *Whiteinella baltica*, (*7*) *Wh. archaeocretacea*, (*8*) *Wh. aumalensis*, (*9*) *Wh. holzli*, (*10*) *Wh. paradubia*, (*11*) *Dicarinella hagni*, (*12*) *Praeglobotruncana oraviensis*, (*13*) *Pr. imbricata*, (*14*) *Pr.* cf. *praehelvetica*; II, benthic foraminifers: (*15*) *Gaudryina filiformis*, (*16*) *Arenobulimina* spp., (*17*) *Lenticulina* spp., (*18*) *Brotzenalla berthelini*, (*19*) *Lingulogavelinella globosa*, (*20*) *Gyroidina turgida*; III, planktonic to benthic form ratios. Key 1, 2, planktonic foraminifers (1, ratites, 2, keeled taxa); 3, benthic foraminifers; 4, localites of samples; 5, limestones; 6, bituminous shales; 7, silty shales with bioturbation marks.

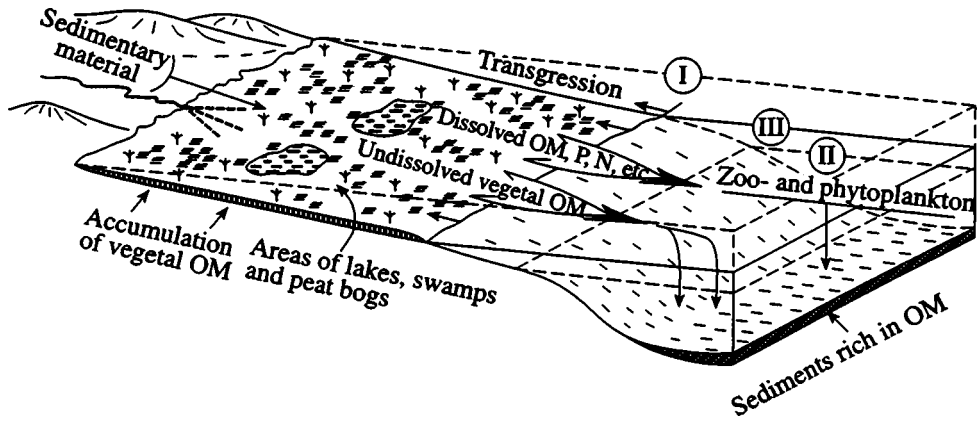

Fig. 3. Schematic model of interaction between the transgressing sea and coastal lacustrine–swamp landscapes (after Gavrilov, 1994). I–III. Position of sea level: I, before regression, II, during maximal regression, III, during developing transgression.

A comparison of the Mesozoic–Cenozoic eustatic oscillation curve (Haq *et al.*, 1987) with the distribution of deposits containing OM shows that not all transgressions caused the formation of black shales. Obviously, certain conditions should have been present to somehow regulate this kind of sedimentation. Normally, the transgressions, associated with the formation of horizons rich in OM, were preceded by more or less developed regressive episodes that initiated specific paleogeographic environments. The regressions exposed great areas of sea floor in relatively shallow epicontinental seas or on the shelfs of basins (Fig. 3). Geomorphologically, they were lowlands flattened out by sea erosion and sedimentation. In predominantly warm and humid climates, the vast newly formed coastal plains were favorable for the development of the limnetic–swamp and adjoining lagoon landscapes. The former probably appeared very quickly: as the data on recent peat accumulations indicate, the impressive peat bogs, many meters thick, were formed during several hundreds or a few thousand years (Blaine, 1989; *et al.*). Therefore, the regressions, initiated by eustatic fluctuations, caused the appearance of vast swamped territories on the earth's surface, so that the sea basins were banded on their perimeters by coastal plains, where considerable amounts of vegetal OM accumulated.

After a certain time period, however, at least in the case of the short-period eustatic fluctuations, it was usually after a few tens of thousands of years that a regressive stage changed to a rapidly developing transgression. During this latter stage, the advancing sea actively interacted with the coastal limnetic–swamp landscapes. The process of erosion of the accumulated vegetal OM into the sea basins was not limited by the mechanical transportation from the coastal landscapes. In addition to the solid OM, the swampy basins stored large amounts of the dissolved OM, the product of disintegration of the available OM mass. Moreover, the process of OM accumulation and partial dissolution was persistently maintained during the entire time of existence of these landscapes. The transgressing sea advanced over the swamped basins and reduced the run-off distance of the drainage waters, thus providing an easy supply of dissolved OM into the sea basin, where it was partially buried in sediments and increased their C_{org} content; the other part, most likely the dissolved OM, was involved in a new biological cycle being utilized by plankton, thus raising its bioproductivity (Fig. 3).

An analysis of mineralogical–geochemical environments in modern swamp systems indicates that the process of peat formation creates conditions favorable for the chemical activity of phosphorus, an important element in the biological cycle. The swamp evolution process, on the whole, can be regarded as one of the typical means of geochemical migration of phosphorus in the hypergenesis zones (Kovalev, 1985). Even if the territory was not swamped, the accumulations and deposits of phosphates were actually formed under hypergenic weathering, especially in the tropical areas. For example, the phosphate eluvium is one of the frequent formations in outcrops of the Cretaceous carbonate deposits in central Russia (Mikhailov, 1986). We, therefore, suggest that the interaction between the sea and the coastal landscapes provided the means for carrying off significant amounts of phosphorus compounds into the sea basin. A supply of these substances to the sea could have greatly increased its bioproductivity and, given certain conditions, could have enriched its sediments with OM.

The aggressive environment of swampy systems was favorable for the reworking of alluvial sediments, when these systems interacted with the underlying rocks. Quite a number of terrigenous minerals, normally known to be resistant to weathering, were dissolved in a peat bog environment and enriched waters with many microelements. These latter were carried off by the drainage network to the water basins and were involved into the biological cycle. Therefore, the peat bogs served as a sort of filter and as a reworking system

of the sedimentary material supplied from the inner parts of the continents. It should also be taken into account that the coastal lacustrine–swamp landscapes primarily developed not on ancient rocks, but also on poorly lithified or nonlithified deposits, the former shelf sediments. That is why they were so easily affected by reworking in the aggressive peat bog environment.

Accordingly, we believe that the interaction between the advancing sea and the lacustrine–swamp landscapes that appeared on the coastal plains as a result of the previous regression was the mechanism, on the one hand, supplying the water basins with considerable amounts of vegetal material and, on the other hand, promoting the flourishing of zoo- and phytoplakton mainly owing to the run-off of the dissolved biophilic elements into the water basin (Gavrilov, 1994). This mechanism was responsible for the formation of rock horizons enriched in OM of either vegetal (land) or planktonogenic origin in accordance with the conditions of sedimentation in that, or another setting in the basin; as a rule, the two processes were concurrent.

Furthermore, the lacustrine–swamp landscapes, particularly the peat bogs, are rather unstable systems, and the wave erosion during transgression often completely demolished them, leaving almost no trace in the geological records except in the regions with an active coal accumulation and specific tectonic regime. That is why the sedimentary sections rarely show direct marks of the interaction between marine and continental coastal landscapes. There are certain indirect evidence, among them geochemical (Gavrilov, 1992), that seem to generally confirm our conclusion. The observation data on artificial reservoirs during their filling also imply this, because such basins are miniature models of a developing transgression. It was noted, that during the first years of filling, an extremely intensive burst of phytoplankton productivity was caused by the growth of concentration of biogenic elements washed out of the inundated soils (Petrova, 1990; Tseeb, 1962). In essence, this phenomenon is an analogue to the processes that took place during the full-scale sea transgressions.

We should bear in mind, though, that providing the supply of biogenic matter and raising the productivity of the basins, the interaction of sea transgressions with the coastal landscapes could be favorable for the formation of OM rich sediments, whereas the realization of this process depended on the combination of different factors. For example, in the basin with a dominant biogenic carbonate sedimentation, the growth of bioproductivity along with preservation of the organic forms traditional for the basin can simply result in the growth of the sedimentation rate, and, in this case, the geological section of sediments will not show any appreciable changes in the character of sedimentation. Consequently, the impression that transgression has no effect on sedimentation can be wrong. A different pattern appeared in cases when the burst of bioproductivity was followed more or less apparent change in the plankton composition and sudden increase of shell-less forms (dinoflagellates, bacterial plankton, etc.); as a result, the sediments were enriched with OM, if the background sedimentation rates were relatively low.

Further, we shall analyze the consequences of such bursts of bioproductivity accompanied by the structural reconstruction of plankton assemblages and accumulation of excessive amounts of OM in sediments.

STRESS EFFECT OF GEO- AND BIOCHEMICAL FACTORS ON BIOTA

Toxic chemical elements. The high OM content in sediments causes intensive diagenetic processes that determine the geochemical environment not only in sediments, but, to a certain extent, in bottom water as well. The abundance of OM in sediments also affects the character of mainly the anaerobic type of diagenesis throughout the entire sedimentary sequence. The absence of the oxidizing film at the sedimentary surface encourages a considerable supply of certain chemical elements and their compounds from the bottom water into sediments (Volkov, 1980; *et al.*). Owing to this and the sufficiently high absorption capacity of the dead plankton tissues at the sedimentation stage, the silts become enriched in many heavy metals: Cu, Pb, Ni, Hg, Co, Mo, Ag, Cd, Zn, As, and others. According to the degree of their toxicity, the elements can be divided into three groups (Wood, 1984): nontoxic (Na, K, Mg, Ca, H, O, N, C, P, Fe, S, Cl, Br, F, Li, Rb, Sr, Ba, Al, Si); highly toxic and easily absorbed by organisms (Be, Co, Ni, Cu, Zn, Sn, As, Se, Te, Pd, Ag, Cd, Pt, Au, Hg, Tl, Pb, Sb, Bi); and toxic, either very rare or occuring in hardly soluble compounds (Ti, Hf, Zr, W, Nb, Ta, Re, Ga, La, Os, Rh, Ir, Ru). Most of the elements accumulating in sediments rich in OM are those of the highly toxic group. The data on the geochemistry of black shales indicate that they generally contain large amounts of toxic trace elements (Yudovich and Ketris, 1991). The toxicity of metals in these groups is different for individual organisms. Somers (1959) compared the relative toxicity of metals for some of the groups of organisms and suggested the following toxicity sequence that corresponds well to the electronegativity values of elements: Hg > Cu > Pb > Pd > Ni > Co > Be > Zn > Mn > Sr > Mg > Ca > Ba. The sapropelic horizons often lack correlation between pyritic sulfur and heavy metals. This implies that heavy metals in sediments occur in forms more mobile than sulfides and may exercise their toxic properties. The diagenetic (as well as sedimentary and biogenic) enrichment of silts with toxic heavy metals has a negative effect on benthic fauna, particularly on the burrowing species.

The studied influence of microelements in the habitat environments on microorganisms (Letunova and Koval'skii, 1978) reveals that high concentrations of

Co, Mo, Cu, V, U, Sc, in most cases, inhibit biota growth. The degree of this influence is different for similar forms in different biochemical provinces. Letunova and Koval'skii remarked that, in a population, the presence of organisms, which are variably adapted to geochemical factors and show an inherited fixed threshold sensitivity to various concentrations of chemical elements, triggers the process of natural selection that becomes more vigorous under extreme conditions.

Hydrosulfuric poisoning of bottom waters. The effect of toxic trace elements on biota is often eclipsed by the action of another product of anaerobic diagenesis, the hydrogen sulfide. The toxic properties of H_2S and its effect on organisms are well known. The hydrosulfuric poisoning of water in basins with more or less stable anoxic environments mainly results from the H_2S migration from OM-rich sediments to bottom waters, as in the Black Sea. The amount of H_2S escaping from sediments depends, however, on the intensity of sulfate reduction and the capacity of sediments to fix the resultant hydrogen sulfide. The latter process is determined by the content of the reacting forms of metals, primarily iron, in sediments. As long as the content of reacting iron is greater than that of H_2S, the released hydrogen sulfide is completely absorbed in the sediments for the formation of Fe sulfide (Volkov, 1984). The dynamics of the process definitely implies that the accumulation of either the terrigenous or carbonate sediments rich in OM (as the extreme members of the continuous clay–limestone series) differently affects the commencement of hydrosulfuric poisoning of a basin. In fact, the average Fe content in clays is almost an order higher than in limestone, i.e., 4.72% versus 0.38%, respectively (Turekian and Wedepohl, 1961). Therefore, if the OM content in deposits is the same, then the clays have a much higher H_2S fixing potential than the carbonate sediments. We should keep in mind, of course, that not all the iron in sediments posesses the same reactive ability and is involved in sulfide formation; however, this does not alter the process in general.

In basins with mostly terrigenous sedimentation, the OM enrichment may be unable to provoke hydrosulfuric poisoning. The process starts only when the OM accumulation becomes much more intensive than the Fe concentration, though somewhat later than the initial enrichment of sediments with OM, when the reacting Fe reserves would run out. On the other hand, the OM enrichment of carbonate sediments with a low Fe content can almost instantly launch the evacuation of diagenetic H_2S from sediments and the hydrosulfuric poisoning. The thickness of the poisoned water layer is primarily controlled by the intensity of diagenetic processes in sediments, the hydrodynamic regime in the basin, and water stratification.

In the Cenomanian–Turonian boundary layers of Crimea (Fig. 2), the sulfur content in rocks with OM is 1.5–2% lower than in the rocks upsection or downsection, i.e., almost twice as less.[1] We believe that this kind of sulfur distribution resulted from intensive reduction processes and evacuation of a large share of generated H_2S from the sediments. The hydrosulfuric poisoning of the basin was one of the principal factors responsible for the distribution of organisms in the section (Fig. 2). The sulfide formation in these OM-rich deposits, however, was insignificant due to their low iron content.

The toxic element accumulation in silts with higher OM and hydrogen sulfide contents was the leading factor that negatively affected the benthic fauna. However, many plankton forms, such as foraminifers, concurrently disappear from the sediments, which accumulated during the anoxic events (Fig. 2). The biological cycle of foraminifers did not necessarily involve sinking to the deeper water levels that could have been poisoned by hydrogen sulfide. For instance, some of the taxa were visibly affected by stress processes in the upper water layers, where mainly the cosmopolitan *Hedbergella* forms were likely to survive. The absence of carbonate shells in black shales could have been partly caused by their diagenetic dissolution. Nevertheless, this circumstance alone cannot account for the actual distribution patterns of fauna in the geological sections. For example, in the Cenomanian–Turonian boundary deposits (Fig. 2), the bituminous marls show a general decline in the carbonate content that, apparently, reflects the reduction of $CaCO_3$ accumulation in sediments caused by the lower productivity of certain planktonic calcareous forms. The rise of the OM content in sediments, however, is the result of the flourishing of carbonateless phytoplankton and, perhaps, also of the bacterial mass. Let us assume, that the formation of an OM-rich horizon was connected not with the planktonic assemblage reorganization, but only with the bioproductivity growth in a basin with an intact former biotic structure; then, as noted above, there would have been a simple increase in sedimentation rate without the growth of OM content in sediments, and no sharp lithological–geochemical differences in the section units.

There is evidence that during the anoxic events, the phytoplankton assemblages essentially changed; thus,

[1] The manuscript was already submitted for publication when the paper by Naidin and Kiyashko (1994) was published. The authors also presented data on the distribution of some chemical elements in the Cenomanian–Turonian boundary units from the Aksu-Dere section in Crimea. Unlike the pattern shown in Fig. 2, these elements show considerable scattering of their concentration values inside the OM-rich layer. We believe that the discrepancies between the results cited and our results are caused by the different way of sampling: the cited authors have, apparently, collected all of the rock varieties conspicuous within the layer, including those from the light-colored lenslike interbeds, where the contents of C_{org} and other elements are much lower. We think that the formation of light-colored lenslike patches inside the black shales was caused by hypergenic processes, i.e., by oxidation of occasional diagenetic pyrite concretion during the weathering. The subsequent action of the resultant sulfuric and sulfanilic acids on rocks changed them, particularly, by bleaching and lowering their $CaCO_3$ content (Perel'man, 1972). For our analyses, we used rocks mainly unaltered by hypergenic processes.

the data on the distribution of palynomorphic groups in the lower Toarcian Posidonia shales shows that some forms of the phytoplankton disappear, whereas other forms are particularly abundant (Wille, 1985; Riegel *et al.*, 1986). Therefore, the accumulation of OM-rich sediments was concurrent with the stress effect on biota, both in the bottom and upper water strata of the basin. However, it is hardly probable that the stress effect was the result of hydrosulfuric poisoning of waters on such a large scale, in some cases almost global.

Red tides and their effect on biota. One of the possible causes of the sudden bioproductivity increase in the surface water layers and of the stunted state, if not complete extinction, of the numerous planktonic forms inhabiting marine basins was the active bloom of phytoplankton, or the so-called red tides. The oceanographic studies of red tides indicate that they are quite common in nature; they are observed off the coasts of nearly all continents and in the inner seas (Tumantseva, 1985; Anderson *et al.*, 1982; Bodeanu and Usurelu, 1979; Correto *et al.*, 1986; Kat, 1983; Maclean, 1979; Murphy *et al.*, 1975; Sukhanova *et al.*, 1988; Watras *et al.*, 1985). The contemporary occurrences of this phenomenon were considered by a number of researchers to account for certain peculiarities in biota development in the seas in the geologic past. Naidin was the most persevering advocate of the relationship between the Late Cretaceous biotic events and red tides (Naidin, 1985, 1986; Naidin *et al.*, 1986).

The effect of red tides on ecology could be drastic (Dale and Yentsch, 1978; Ray and Aldrich, 1965; White, 1981; *et al.*). For example, the red tide off the eastern coast of England, in May 1968, was responsible for the extermination of mollusks, eels, and sea birds, mainly cormorants (Adams *et al.*, 1968); eighty people were poisoned from eating mollusks. This red tide was associated primarily with the dinoflagellate bloom, particularly of the *Gonyaulax tamarensis* Lebour (Robinson, 1968). The zone of toxic dinoflagellates stretched along the coast for almost 500 km, and was more than 10 km wide (Indham *et al.*, 1968). The ecosystem received additional stress from the decay of a mass of dead algae that caused a decrease in the oxygen content of the sea.

The mass mortality of marine fauna, including numerous fish, was periodically recorded in the areas of oceanic upwelling. This phenomenon can be caused by an oxygen deficit and hydrogen sulfide formation (Copenhagen, 1953). In the Walvis Bay, however, fish mortalities almost invariably coincided with red tides and, at the time, the *Gonyaulax tamarensis* Lebour (Pieterse and Van der Post, 1967) was one of the four dominant dinoflagellate species.

The red tides are provoked by blooms of both toxic and nontoxic algae. There are toxic diatoms, and up to 200 dinoflagellate species are toxic (Zernova, 1992; Abbott and White, 1979; Loeblich and Loeblich, 1979; Shimizu, 1978; Steidinger and Baden, 1984); the blue-green algae and bacteria can also be neuro-, hepato-, and, apparently, dermatoxic (Petrova and Chernaenko, 1993; Jackim and Gentile, 1968). Moreover, the emission of toxins by some phytoplankton species under certain conditions becomes a part of more extensive events, for instance, in the interaction of plant assemblages followed by the formation of various chemical compounds and their extraction into the environment (Grodzinskii, 1965; Rice, 1978; Khailov, 1971; Lucas, 1947; Sounders, 1957; Whittaker and Feeny, 1971; *et al.*). Numerous experiments confirmed this kind of phytoplankton interaction (Rice, 1978; *et al.*).

It is important to note, that some of the phytoplankton forms are inherently toxic, whereas other forms extract toxins under adverse conditions; e.g., under low concentrations of biophilic elements in their environment, or by an impaired balance between these elements. The observed development of one of the red tides off the eastern coasts of the North Sea in 1989 (Zernova, 1992) shows that its appearance was caused by a combination of conditions that set in after the run-off of excessive amounts of phosphorus and nitrogen from the continent into the sea. It was established that a certain plankton species, e.g., the flagellate alga *Chrysochromulina polylepis* responsible for the red tide, started emitting the toxin during the early stages of its mass development under low phosphorus and high nitrogen contents. This toxin, at a certain stage of its concentration in the environment, caused the mortality of other plankton forms. There were even cases of large fish kills. When the supply of nutrient was exhausted, the growth of the phytoplankton population ceased, and all the dead biomass sank to the sea bottom, essentially altering the geochemistry of the bottom waters and the habitat conditions of benthic fauna and certain fish species. There are numerous examples of similar effects of red tides on the environment (Aubert, 1990; Brongersma-Sanders, 1957; *et al.*).

The situation described above was caused by a combination of several climatic factors in Europe in 1989 that instigated the run-off of considerable quantities of phosphoric and nitric compounds into the sea, and to a lesser extent, by constant anthropogenic eutrophication. In the geologic past, however, analogous situations could have been created in the periods of rapid sea-level fluctuations, when transgressions carried off excessive amounts of biophilic elements from land into the sea, and thus caused productivity bursts in plankton and toxic phytoplakton. The fine lamination in numerous horizons rich in OM testify to the cyclic character of these surges of bloom. We can be fairly positive that these periods were favorable for the development of phenomena analogous to the recent toxic red tides. The phytoplankton bloom, greatly increasing the biomass and, consequently, saturating the silts with organic matter, stunted or killed the leading plankton forms, including the zooplankton with a carbonate shell that inhabited the basin. The annual occurrences of intensive phy-

Content of chemical components in interbreeds with different content of organic matter*

Sample no.	C_{org}	$CaCO_3$	Fe	Mn	Ti	P	Cr	Ni	V	Cu	Co	Pb	Mo	Ag	Stratigraphic and geographic locality of samples
18/1	0.5	74.1	0.89	0.04	0.15	0.05	40	20	60	47	12	10	1.8	0.05	Kuma Formation (Bartonian), Rubaschai River (Dagestan)
18/2	10.6	7.9	3.66	0.02	0.45	0.13	210	105	400	220	30	30	8.8	2.35	"
25	1.53	88.3	3.91	1.46	0.25	0.20	76	21	48	28	5	5	2.4	0.27	Kuma Formation, Kheu River (Kabardino-Balkariya)
26	7.50	54.0	3.25	0.36	0.30	0.27	240	102	1600	220	10	14	14	0.19	"
9	traces	13.6	3.89	0.30	0.51	0.06	141	112	162	79	38	30	2.3	0.08	Upper Paleocene, Kheu River
11a	8.0	traces	5.95	0.16	0.41	0.31	740	338	1100	172	53	69	23.1	6.50	"
11b	7.2	12.9	6.69	0.16	0.38	0.30	270	356	1700	122	52	32	73	2.70	"

* Contents of C_{org}, $CaCO_3$, Fe, Mn, Ti, P, in wt %; other elements in ppm.

toplankton blooms persisted as long as the ample supply of phosphorus, nitrogen, and other elements was maintained. With the termination of transgression, this process gradually attenuated, the bioproductivity of phytoplankton was reduced, and those forms of organisms that survived or escaped to more beneficial environments during the adverse period returned to the areas of their previous habitat (Fig. 2).

Red tides in the geologic past. An analysis of the causes of the formation of "fish cemeteries" in the Oligocene deposits of Ciscaucasia implies that red tides affected biota in the geologic past as well (Mstislavskii and Kochenov, 1960). The Lower Cretaceous deposits in Denmark contain accumulations of bivalve mollusks, the mass mortality of which was associated with the toxic dinoflagellate bloom (Noe-Nygaard *et al.*, 1987).

The finds of marine fauna mass mortality (fishes, mollusks, and others) in ancient geological sections are few, but the red tides were, apparently, rather frequent. There is no direct evidence of toxic red tides, as the toxins in rocks were not discovered, as can now be done, by studying the forms responsible for the bloom in water basins (Carmichael *et al.*, 1993). Moreover, it is not yet clear, whether the toxins can be stored in ancient deposits, at least after the diagenesis stage. Probably, in the course of evolution, the susceptibility of organisms to toxins changed, and the substances that were then poisonous to them are now harmless. Therefore, to establish connections between the formation of certain layers and the toxic red tides, we have to rely on indirect evidence. A sudden replacement of organisms traditional to the basin by other organisms, or even considerable changes in their relative quantities may serve as such evidence. The bursts of dinoflagellate, blue–green algae or, probably, bacteria blooms enrich the sediments with OM, and noticeably alter the geochemistry of deposits. The occurrences of toxic diatom blooms are evidently specifically recorded in the sediments.

In southern Russia and the adjoining areas, the Mesozoic and Cenozoic deposits, at different stratigraphic levels, have layers of different thickness and extent that can be associated with the red tide events.

The middle Eocene marls of the Kuma Formation in Ciscaucasia show a higher background C_{org} content than that in the upsection or downsection deposits. The formation lacks benthic fauna, probably owing to the H_2S poisoning of the bottom waters, but the plankton forms, foraminifers included, are abundant and constitute one of the groups of rock-forming organisms. These deposits have thin intercalations (less than one to few cm thick), with considerably higher C_{org} and reduced $CaCO_3$ contents. This is evident from the comparison of data of chemical analyses of sample pairs from these intercalations (table, Samples 18/2, 26) and a few decimeters apart from them (Samples 18/1, 25). One of the interbeds is particularly interesting (Sample 18/2), because the C_{org} content in it grows from 0.5 to 10%, while the $CaCO_3$ amount is reduced from 74 to 8%. The interbeds either completely lack foraminifers, or contain very rare specimens. Concurrently, as shown in the table, the geochemical features of the deposits are largely changed. The dissolution was not responsible for the lack of foraminifers, because the same interbeds have large amounts of well-preserved coccoliths without a trace of dissolution. These kinds of intercalations were formed during a vigorous reconstruction of the ecosystem in the upper water layer caused, as we believe, by a burst of toxic phytoplankton bloom that stunted the other plankton forms and foraminifers. In the Kumian time, the red tide periods were short and, evidently, affected rather small areas, because as soon as they terminated, the traditional plankton of the basin recovered its habitat.

This problem also concerns the upper Paleocene sapropelic horizon stretching over a large territory, from the Caucasus to Central Asia (Muzylev *et al.*, 1989); it was formed under conditions of a rather short and rapid transgression. In Ciscaucasia (Kabardino–Balkariya), the base of the inhomogeneous sapropelic layer, about 0.4–0.45 m thick, includes a 4- to 5-cm-thick band with C_{org} content up to 8%; the band lacks $CaCO_3$ (table, Sample 11a) and calcareous plankton remains. In the middle of the horizon, the plankton remains appear, the carbonate content rises (table, Sample 11b) (Gavrilov and Muzylev, 1991), and the C_{org} content is still rather high. Among the two beds of the sapropelic horizon with similar C_{org} contents, one contains intact calcareous shells, while they are absent in the other; this indicates that the differences in the degrees of diagenetic reworking of the sediments were not the main cause of shell distribution, and that these differences appeared at the sedimentation stage. We consider it realistic to associate the formation of the interbed at the base of the sapropelic layer with the toxic phytoplankton bloom. This latter process is responsible for the absence of the coccolithoforid and foraminifer skeletons in the sediments. In a later period, during the accumulation of the rest of the horizon, the environment in the basin, perhaps, somewhat changed, the phytoplankton bloom ceased to encourage toxin extraction, and other plankton forms, including zooplankton, returned to this part of the basin and participated in the formation of sapropelic sediments. An intensive phytoplankton bloom at the end of the late Paleocene developed concurrently with the rapid global transgression and, evidently, was one of the forms of biota reaction to the changing environment in the basins; in other regions, this event could have occurred and been recorded in sediments differently. It is obvious, however, that in most localities, the stress effect of the bloom on the marine ecosystem was dramatic. Thus, according to Muzylev's data (Gavrilov and Muzilev, 1991), the boundary between the nannoplankton subzones lies inside the sapropelic layer, as evident from nannoplankton distribution.

An analysis of Cretaceous fauna reveals that in many localities, the Cenomanian–Turonian anoxic event was associated with the formation of C_{org}-rich deposits, and was followed by the extinction of highly specific "deep-water" taxa. To complete their life cycle, they must have sunk to the lower water layers. The hydrosulfuric poisoning of the bottom waters drastically reduced the reproductive processes in the euphotic zone. In a number of cases, however, the "shallow-water" taxa, whose life cycle is confined to the euphotic zone, disappeared from the geological units. These taxa seem to exemplify another type of selection with numerous quickly maturing offspring and a high competition between the young and adult specimens (r-selection). The regulating function for this group of fossil organisms was exercised, evidently, by the processes in the euphotic zone itself. Leary *et al.* (1989) suggested a rather attractive scheme, though not always accurate in its second part (C, D, and E). The mortality of planktonic foraminifers of the *Hedbergella* and *Whiteinella* genera was, apparently, caused by stress events in the surface water layers. The most resistant to this process were the small forms of cosmopolitan taxa with little shells and globular chambers of the *Hedbergella* type. The oligotaxic period in the development of planktonic foraminifers of the Globotruncanidae family coincided with the period of the highest level of the World Ocean. The variations in the oceanographic parameters could have acted as powerful controlling factors in the phylogenetic development of foraminifers; e.g., a deficit in O_2 in the bottom waters often caused the appearance of H_2S and concurrent considerable amounts of toxins in the euphotic zone as a result of bacterial or phytoplankton blooms. These events not only destroyed the highly specific morphotypes of the *Rotalipora* and *Thalmanninella* genera, but also caused the formation of new morphological characteristics promoting higher degrees of specialization of organisms (Fig. 1).

In this respect, the Cretaceous–Paleogene boundary is of particular interest. In some regions, the Maastrichtian–Danian boundary corresponds to a small hiatus and, in other regions, to a clayey layer of few decimeters thick, often brown and located between carbonate rocks upsection and downsection. In some localities, the Cretaceous–Paleogene boundary deposits were formed in anoxic environments (Kajiwara and Kaiho, 1992). The sea-level fluctuations, the anoxic environments, and a sudden decline in the productivity of calcareous plankton were also typical of the Cenomanian–Turonian and late Paleocene events mentioned above. It is quite probable, that during the Maastrichtian–Danian event, the calcareous plankton suffered strong stresses due to the development of toxic red tides that could have led to biotic crises in a number of regions of the World Ocean. In several publications, Naidin suggests that red tides were one of the causes of the biotic crises at the Cretaceous–Paleogene boundary (Naidin, 1985, 1986; Naidin *et al.*, 1986).

The opponents of the idea that red tides essentially influence biotic life usually refer to the fact that the modern red tides are local and brief and, therefore, cannot cause global events. We should bear in mind, however, that environments in the past geological epochs could be quite different from those we have today. Thus, during the Cretaceous, there was the so-called "sluggish ocean," with an extremely slow circulation and lacking active contour currents. In this ocean, the red tides, stimulated by a developing transgression, could cover vast territories and, once started, could continue for a rather long period. Consequently, they could cause the extinction of certain specific groups of organisms, and further, the rapid expansion of other groups.

The problem of red tides in the geological past, their effect on biota, and the definition of criteria for their identification are, as yet, insufficiently elaborated. But this phenomenon occurs in our time, and we can hardly overlook the fact that it could have occurred in ancient epochs, though in a somewhat altered form. We assign red tides, or similar phenomena, to factors that produce the strongest stress effect on biota, both in marine and freshwater basins, and, therefore, should allow them due consideration in paleoecological and lithological reconstructions.

CONCLUSION

In sum, the scheme of biota evolution under factors responsible for the accumulation of black shales can be as follows. Rapid sea-level fluctuations created conditions for a more intensive supply of higher amounts of biophilic elements from land to water basins and caused sudden bursts of plankton productivity, particularly the shell-less phytoplankton. Under certain conditions, e.g., in the case of an unbalanced supply of nutrient, certain phytoplankton species secreted toxins that stunted or even killed other forms of organisms, in particular, the calcareous plankton. As a consequence, the sediments accumulated OM derived from shell-less algae and other organisms, and were concurrently depleted in $CaCO_3$ due to a smaller share of foraminifera, coccolithophorids, and other carbonate fossils in the sediments.

The abundance of OM in silts caused an intensive development of diagenetic processes, which (a) were primarily responsible for the accumulation of numerous toxic heavy metals in silts, and (b) generated a large amount of H_2S. Both the toxic elements and H_2S had a negative effect on bottom organisms. Under the conditions of anaerobic diagenesis, the H_2S migration was largely controlled by the intensity of processes of diagenetic sulfide formation. Provided that all other factors are equal, the basins where sediments had a lower Fe content were more favorable for hydrosulfuric poisoning than the basins that accumulated sediments relatively rich in Fe. Since in carbonate sediments the Fe content, as a rule, is much lower than in clayey sediments, the hydrosulfuric poisoning of water in the basins with carbonate sedimentation, apparently, developed more intensively than in those with terrigenous sedimentation.

Therefore, as the OM-rich horizons accumulated, different factors, including geo- and biochemical, affected both the benthic and planktonic biota. The mechanisms of these stresses were different, but the results were similar: throughout the basin system, from its bottom to surface layers, the conditions adverse to the existence of organisms could be generated more or less synchronously. This combined stress on biota by the stated factors could take place together with other effects that accompanied the sharp oscillations of the ocean level, such as the changes in the area, salinity (?) and temperature of a basin, the greater stratification of its waters, etc. All of these factors were obviously essential to biota reorganization, i.e., the extinction of the old and the flourishing of the new forms.

REFERENCES

Abbott, B. and White, A., Toxigenesis in Dinoflagellates, *Toxic Dinoflagellate Blooms,* Taylor, D.L. and Seliger, H.H., Eds., New York: Elsevier, 1979, pp. 494–496.

Adams, J.A., Seaton, D.D., Buchanan, J.B., and Longbottom, M.R., Biological Observation Associated with the Toxic Phytoplankton Bloom off the East Coast, *Nature,* 1968, vol. 220, no. 5162, pp. 24–25.

Anderson, D.M., Kulis, D.M., Orphanos, J.A., and Ceurvels, A.R., Distribution of the Toxic Dinoflagellate *Gonyaulax tamarensis* in the Southern New England Region, *Estuarine, Coastal Shelf Sci.,* 1982, vol. 14, pp. 447–458.

Arthur, M.A., Brumsack, H.-J., Jenkyns, H.C., and Schlanger, S.O., Stratigraphy, Geochemistry and Paleoceanography of Organic Carbon-rich Cretaceous Sequences, *Cretaceous Resources, Events and Rhythms*, Ginsburg, R.N. and Beaudoin, B., Eds., NATO ASI Series, Series C, vol. 304, Dordrecht: Klywer Acad., 1990, pp. 75–119.

Aubert, M., La Proliferation Estivale du Plancton Toxique, *La Recherche,* 1990, vol. 21, no. 223, pp. 916–919.

Blaine, C.C., Modern Analogs of Coal Formations, *U.S. Geol. Surv. Yearb., Fiscal Year 1988, Denver (Col.),* 1989, pp. 81–84.

Bodeanu, N. and Usurelu, M., Dinoflagellate Blooms in Rumanian Black Sea Coastal Waters, *Toxic Dinoflagellate Blooms,* Taylor, D.L. and Seliger, H.H., Eds., New York: Elsevier, 1979, pp. 151–154.

Brockamp, B., Zur Entstehung Deutscher Eisenerzlagerstaetten, *Arch. Lagerstaettenforsch.,* Berlin, 1942, no. 75, pp. 7–86.

Brongersma-Sanders, M., Mass Mortality in the Sea: Treatise on Ecology and Paleoecology, vol. 2, *Mem. Geol. Soc. Am.,* 1957, vol. 67, pp. 941–1010.

Brumsack, H.-J., Trace Metal Accumulation in Black Shales from the Cenomanian/Turonian Boundary Event, *Global Bio-Events, Lecture Notes in Earth Sciences, vol. 8,* Walliser, O., Ed., Heidelberg: Springer, 1986, pp. 337–343.

Carmichael, W.W., Chernaenko, V.M., and Evans, W., Cyclic Peptide Hepatotoxins of the Fresh Water Cyanobacteria (Blue–Green Algae) from Blooming Water Basins of Ukraine and European Russia, *Dokl. Ross. Akad. Nauk,* 1993, vol. 330, no. 5, pp. 661–663.

Copenhagen, W.J., The Periodic Mortality of Fish in the Walvis Region, *Invest. Rep. Div. Fish. S. Afr.,* 1953, vol. 14, pp. 1–35.

Correto, J., Benavides, H., Negri, R., and Glorioso, P., Toxic Red-Tide in the Argentine Sea: Phytoplankton Dinoflagellate *Coniaulax excavata* in a Frontal Area, *J. Plankton Res.,* 1986, vol. 8, pp. 15–28.

Dale, B. and Yentsch, C.M., Red Tide and Paralytic Shellfish Poisoning, *Oceanus,* 1978, vol. 21, pp. 41–49.

Gavrilov, Yu.O., Geochemistry of Terrigenous Deposits as Regards Eustatic Oscillations of the Sea Level: the Lower and Middle Jurassic, Ciscaucasia, *Litol. Polezn. Iskop.,* 1992, no. 2, pp. 57–70.

Gavrilov, Yu.O., Possible Causes of Deposit Accumulation Rich in Organic Matter Due to Eustatic Oscillations of Sea Level, in *Problemy Evolutsii Biosphery* (The Problems of Evolution of the Biosphere), Moscow: Nedra, 1994, pp. 305–311.

Gavrilov, Yu.O. and Muzylev, N.G., Geochemistry of Sapropelic Interbeds in Paleogene Deposits of the Central Caucasus, *Litol. Polezn. Iskop.*, 1991, no. 6, pp. 84–98.

Gorbachik, T.N. and Kopaevich, L.F., *Vliyanie melovykh sobytii na evolyutsiyu globigerinid,* Effect of Cretaceous Events on the Evolution of Globigerinids, *Geologicheskaya istoriya Arktiki v Mezozoe i Kainozoe* (Geological History of the Arctic in the Mesozoic and Cenozoic), St. Petersburg: VNIIOkeanogeologiya, 1992, pp. 17–31.

Grodzinskii, A.M., *Allelopatiya v zhizni rastenii i ikh soobshchestv* (Allelopathy in the Life of Plants and Their Assemblages), Kiev: Naukova Dumka, 1965.

Hallam, A. and Bradshaw, M.I., Bituminous Shales and Oolitic Ironstones as Indicators of Transgressions and Regressions, *J. Geol. Sci.*, 1979, vol. 136, pt. 2, pp. 157–164.

Haq, B.U., Hardenbol, I., and Vail, P.R., Chronology of Fluctuating Sea Levels since the Triassic, *Science*, 1987, no. 4793, pp. 1156–1167.

Ingham, H.R., Mason, J., and Wood, P.C., Distribution of Toxin in Molluscan Shellfish Following the Occurrence of Mussel Toxicity in North–East England, *Nature,* 1968, vol. 220, no. 5162, pp. 25–27.

Jackim, E. and Gentile, J., Toxins of a Blue–Green Algae: Similarity to Saxitoxin, *Science,* 1968, vol. 162, pp. 915–916.

Jenkyns, H.C., The Early Toarcian (Jurassic) Anoxic Event: Stratigraphic, Sedimentary and Geochemical Evidence, *Am. J. Sci.,* 1988, vol. 288, pp. 101–151.

Kajiwara, Y. and Kaiho, K., Oceanic Anoxia at the Cretaceous/Tertiary Boundary Supported by the Sulfur Isotope Record, *Palaeogeogr., Palaeoclimatol., Palaeoecol.,* 1992, vol. 99, pp. 151–162.

Kat, M., *Dinophysis acuminata* Blooms in the Dutch Coastal Area Related to Diarrhetic Mussel Poisoning in the Dutch Waddensea, *Sarsia,* 1983, vol. 68, pp. 81–84.

Kauffman, E.G., The Fabric of Cretaceous Marine Extinctions, *Catastrophes and Earth History: The New Uniformitarianism,* Berggren, W.A. and Van Couvering, J.A., Eds., Princeton: Princeton Univ., 1984, pp. 151–237.

Khailov, K.M., *Ekologicheskii metabolizm v more* (The Ecological Metabolism in the Sea), Kiev: Naukova Dumka, 1971.

Kovalev, V.A., *Bolotnye mineralo–geokhimicheskie sistemy* (Swamp Mineralogical–Geochemical Systems), Minsk: Nauka i Tekhnika, 1985.

Leary, P.N., Carson, G.A., Cooper, M.K., *et al.*, The Biotic Response to the Late Cenomanian Oceanic Anoxic Event: Integrated Evidence from Dover, Southeast England, *J. Geol. Soc.,* 1989, vol. 146, pt. 2, pp. 311–317.

Letunova, S.V. and Koval'skii, V.V., *Geokhimicheskaya ekologiya microorganizmov* (Geochemical Ecology of Microorganisms), Moscow: Nauka, 1978.

Loeblich, A.R. III and Loeblich, L.A., The Systematics of *Gonyaulax* with Special Reference to Toxic Species, *Toxic Dinoflagellate Blooms*, Taylor, D.L. and Seliger, H.H., Eds., New York: Elsevier, 1979, pp. 41–46.

Lucas, C.E., The Ecological Effects of External Metabolites, *Biol. Rev.,* 1947, vol. 22, no. 3, pp. 270–295.

Maclean, J.L., Indo-Pacific Red Tide, *Toxic Dinoflagellate Blooms,* Taylor, D.C. and Seliger, H.H., Eds., New York: Elsevier, 1979, pp. 173–178.

Mikhailov, B.M., *Rudonosnye kory vyvetrivaniya* (Ore-bearing Weathering Crusts), Leningrad: Nedra, 1986.

Mstislavskii, M.M. and Kochenov, A.V., The Maikopian Bone Breccia and Fish Mortality in "Red Tides," *Dokl. Akad. Nauk SSSR,* 1960, vol. 134, no. 5, pp. 1169–1172.

Murphy, E.B., Steidinger, K.A., Roberts, B.S., *et al.*, An Explanation for the Florida East Coast *Gymnodinium breve* Red Tide of November, 1972, *Limnol. Oceanogr.,* 1975, vol. 20, pp. 481–486.

Muzylev, N.G., Ben'yamovskii, V.N., and Tabachnikov, I.P., Sapropelic Interbeds in the Lower Paleogene Deposits in the Southern USSR, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1989, no. 11, pp. 117–119.

Naidin, D.D., The Cretaceous–Paleogene Boundary: Stratigraphic Aspects, *Bull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1985, vol. 60, issue 5, pp. 71–85.

Naidin, D.P., The Cretaceous–Paleogene Boundary in Mangyshlak and Probable Events at the Maastrichtian–Danian Boundary, *Izv. Vyssh. Uchebn. Zaved., Geol. Razved.,* 1986, no. 9, pp. 3–13.

Naidin, D.P., Pokhialainen, V.P., Kats, Yu.I., and Krasilov, V.A., *Melovoi period. Paleogeografiya i paleookeanologiya* (The Cretaceous Period: Paleogeography and Paleoceanology), Moscow: Nauka, 1986.

Naidin, D.P. and Kiyashko, S.I., Geochemical Characteristics of the Cenomanian–Turonian Boundary Deposits in the Crimean Mountains. Paper 1: Lithological Composition, Content of Organic Carbon, and Some Other Elements, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.,* 1994, vol. 69, issue 1, pp. 28–42.

Noe-Nygaard, N., Surlyk, F., and Piasecki, S., Bivalve Mass Mortality Caused by Toxic Dinoflagellate Blooms in a Berriasian–Valanginian Lagoon, Bornholm, Denmark, *Palaios,* 1987, vol. 2, no. 3, pp. 263–273.

Perel'man, A.I., *Geokhimiya elementov v zone gipergeneza* (The Geochemistry of Elements in the Hypergenesis Zone), Moscow: Nedra, 1972.

Petrova, N.A., *Subtsessiya fitoplanktona pri antropogennom evtrofirovanii bol'shikh ozer* (Succession of Phytoplankton during Anthropogenic Eutrophication of Large Lakes), Leningrad: Nauka, 1990.

Petrova, N.A. and Chernaenko, V.M., Blue–Green Algae in Freshwater Basins, *Priroda,* 1993, no. 8, pp. 3–9.

Pieterse, F. and Van der Post, D.C., Oceanographical Conditions Associated With Red Tides and Fish Mortalites in the Walvis Bay Region, *Invest. Rep. Mar. Res. Lab. S. W. Afr.*, 1967, vol. 14, pp. 1–125.

Ray, S.M. and Aldrich, D.V., *Gymnodinium breve*: Induction of Shellfish Poisoning in Chicks, *Science,* 1965, vol. 148, pp. 1748–1749.

Rice, E.L., *Allelopathy,* New York: Academic, 1974.

Riegel, W., Loh, H., Maul, B., and Prauss, M., Effects and Causes in a Black Shale Event: The Toarcian Posidonia Shale of NW Germany, *Global Bio-Events, Lecture Notes in Earth Sciences, vol. 8,* Walliser, O., Ed., Berlin: Springer, 1986, pp. 267–276.

Robinson, G.A., Distribution of *Gonyaulax tamarensis* Lebour in the Western North Sea in April, May and June 1968, *Nature*, 1968, vol. 220, no. 5162, pp. 22–23.

Shimizu, Y., Dinoflagellate Toxins, *Marine Natural Products* Scheuer, P.J., Ed., New York: Academic, 1978, vol. 1, pp. 1–43.

Simoneit, B.R.T., Organic Geochemistry of Black Shales from the Deep Sea Drilling Project: A Summary of Occurrences from the Pleistocene to the Jurassic, *Mitt. Geol.-Paleont. Inst., Univ. Hamburg*, 1986, no. 60, pp. 275–309.

Somers, E., Fungitoxicity of Metal Ions, *Nature*, 1959, vol. 184, no. 4684, pp. 475–476.

Sounders, G.W., Interrelations of Dissolved Organic Matter and Phytoplankton, *Bot. Rev.*, 1957, vol. 23, pp. 389–409.

Steidinger, K.A. and Baden, D.G., Toxic Marine Dinoflagellates, *Dinoflagellates*, Spector, D.L., Ed., New York: Academic, 1984, pp. 201–261.

Strakhov, N.M., *Zhelezorudnye fatsii i ikh analogi v istorii Zemli (opyt istoriko-geologicheskogo analiza protsessa osadkoobrazovaniya)* (Iron-Ore Facies and Their Analogues in the History of the Earth: An Experience of Historical–Geological Analysis of the Sedimentation Process), Moscow: Akad. Nauk SSSR, 1947.

Sukhanova, I.N., Flint, M.V., Hibaum, G., *et al.*, *Exuviaella cordata* Red Tide in Bulgaria Coastal Waters (May to June 1986), *Mar. Biol.*, 1988, vol. 99, no.1, pp. 1–8.

Tseeb, Ya.Ya., Certain General Regularities in the Formation of the Hydrobiological Regime in the Kakhovka Water Reservoir, *Trudy zonal'nogo soveshchaniya po tipologicheskomu i biologicheskomu obosnovaniyu rybokhozyaistvennogo ispol'zovaniya vodoemov yuzhnoi zony SSSR* (Proc. Zonal Meeting on Typological and Biological Substantiation of Fish Production Management in the Water Basins of the Southern Zone of the USSR), Kishinev, 1962, pp. 204–210.

Tumantseva, N.I., "Red Tide" in the Black Sea, *Okeanologiya*, 1985, vol. 25, issue I, pp. 130–132.

Turekian, K.K. and Wedepohl, K.H., Distribution of Elements in Some Major Units of the Earth's Crust, *Bull. Geol. Soc. Am.*, 1961, vol. 72, no. 2, pp. 175–192.

Volkov, I.I., *Geokhimiya sery v osadkakh okeana* (Geochemistry of Sulfur in Oceanic Sediments), Moscow: Nauka, 1984.

Volkov, I.I., Redistribution of Chemical Elements during Sediment Diagenesis, in *Geokhimiya diageneza osadkov Tikhogo okeana (Transokeanskii profil')* (Geochemistry of Sediment Diagenesis in the Pacific Ocean: A Transoceanic Profile), Moscow: Nauka, 1980, pp. 144–168.

Watras, C., Garson, V., Olson, R., *et al.*, The Effect of Zooplankton Grazing on Estuarine Blooms of the Toxic Dinoflagellate *Gonyaulax tamarensis*, *J. Plankton Res.*, 1985, vol. 7, pp. 891–908.

White, A.W., Marine Zooplankton Can Accumulate and Retain Dinoflagellate Toxin and Cause Fish Kills, *Limnol. Oceanogr.*, 1981, vol. 26, pp. 103–109.

Whittaker, R.H. and Feeny, P.P., Allelochemics: Chemical Interactions between Species, *Science*, 1971, vol. 171, no. 3973, pp. 757–770.

Wille, W., *Palynology of Upper Liassic Bituminous Shales: Cyclic and Event Sedimentation*, Einsele, G. and Seilacher, A., Eds., Berlin: Springer, 1982.

Wood, J.M., Metabolic Cycles for Toxic Elements in the Environment, *Science*, 1984, pp. 1049–1052.

Yudovich, Ya.E. and Ketris, M.P., *Geokhimiya i rudogenez toksichnykh elementov–primesei (Cd, Hg, As, Sb, Se) v chernykh slantsakh* (Geochemistry and Ore Genesis of Toxic Trace Elements (Cd, Hg, As, Sb, Se) in Black Shales), Syktyvkar, 1991.

Zernova, V.V., The Ocean Hits Back with Physiological Weapons, *Priroda*, 1992, no. 5, pp. 38–41.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 326–340. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 15–30.
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Allochthonous Assemblages of Paleozoic Benthic Fossils from the Caspian Syneclise

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Received May 5, 1994

Abstract—The peculiarities of the allochthonous assemblages of Caspian Paleozoic benthic fossils, which form separate allochthonous communities or represent components of mixed complexes, are discussed. Allochthonous assemblages were studied by paleontologic, paleoecologic, and sedimentologic methods. Two types of allochthonous benthic assemblages, namely, synchronous and asynchronous with enclosing sediments, were distinguished. The distinguished types differ mainly in terms of paleontologic criteria. The distribution of allochthonous assemblages depends on the age and facies composition of deposits. Allochthonous assemblages were recorded in the Lower Devonian, Lower Tournaisian, Lower and Upper Visean, Serpukhovian, Lower Bashkirian, Asselian, Sakmarian, and Artinskian sediments. Allochthonous synchronous assemblages are closely associated with sedimentation paleoenvironments and occur in lagoon facies, wave zone shoals, organogenic buildups, and submarine slope, submarine fans, open shelf, and basin facies. Heterochronous allochthonous assemblages are most widely distributed at the base of transgressive cycles.

Key words: *benthic fossils, asynchronous and synchronous assemblages, distinguishing criteria, facies confinement, Caspian syneclise.*

INTRODUCTION

Allochthonous assemblages of benthic fossils were recorded in the Paleozoic of the southern Urals (Rozovskaya, 1952; Rauzer- Chernousova, 1965) and Middle Asia (Poyarkov, 1973; Dzhenchuraeva and Soloshenko, 1983); in the Carboniferous (Klenina and Zaitseva, 1995), Upper Carboniferous (Goroshkova *et al.*, 1989), and Lower Permian (Izotova, 1985; Kukhtinov *et al.*, 1982; Zolotukhina *et al.*, 1989; Zolotukhina and Dan'shina, 1992; Kukhtinova and Kukhtinov, 1990) of the Caspian syneclise; and in the Lower Permian of the west Mugodzhary region (Zamarenov, 1970).

Recent investigations of benthic fossils from the Devonian, Carboniferous, and Lower Permian sediments, recovered by boreholes on flanks of the Caspian syneclise and in adjacent territories, revealed the autochthonous character of most localities. Benthic organisms, namely, foraminifers, corals, bivalve mollusks, bryozoans, brachiopods, crinoids, and calcareous algae, were mainly autochthonously or parautochthonously buried. Allochthonous and mixed assemblages are rarer, though the formers are of prime interest, as they present difficulties in the exact age determination of sediments and are good indicators of sedimentation paleoenvironments.

Allochthonous assemblages were studied according to the following major lines: (1) development of diagnostic criteria of allochthonous assemblages, (2) elucidation of stratigraphic distribution, (3) analysis of the facies control, and (4) study of the ways of their formation.

The available material, together with detailed biostratigraphic and sedimentologic data, enabled the distinction of two types of allochthonous benthic assemblages, synchronous and asynchronous with enclosing sediments, characterized by specific features, vertical and lateral distribution, and by the ways of their formation.

MATERIAL

Allochthonous assemblages were recorded at different stratigraphic levels in all marginal parts of the Caspian syneclise (Fig. 1). The oldest allochthonous benthic fossils were found in Devonian sediments recovered by Well D-2 from the Karachaganak area; at several stratigraphic levels of Carboniferous sediments in Well 6 from the Noven'kaya area, Wells 1 and 2 from the East Vetelkino area, Well 2 from the Burli area, Well P-41 from the Pervosovetsk area, Well P-3 from the Rozhki area, Well 1 from the Dar'ya area, Wells 7, 13, and 21 from the northern flank zone of the Karachaganak mineral deposit, Wells 27 and 32 from the Astrakhan mineral deposit, Well 200 from the Nikolaev area, Wells 2 and 3 from the southwestern flank zone of the Plodovitoe area, Well 1 from Urtatau-Sarybulak, Well 44 from the eastern flank zone of the Tengiz mineral deposit, and in the Lower Permian sediments from the northern (Wells 5, 10, and 11 in the Karachaganak

area and Well P-2 in the Aksai area) and eastern (Well G-1 South Tuskum, Well G-107 Kenkiyak, and Well G-2 East Tobusken) flank zones of the Caspian syneclise.

Allochthonous benthic assemblages include the remains of foraminifers, corals, bivalve mollusks, bryozoans, brachiopods, crinoids, and calcareous algae. Along with the proper allochthonous assemblages, which lack autochthonous elements, mixed complexes with both allochthonous and autochthonous components occur. The revealed mixed assemblages are characterized by different proportions of autochthonous and allochthonous components, though the latter largely dominates.

CRITERIA FOR DISTINGUISHING THE ALLOCHTHONOUS ASSEMBLAGES

Allochthonous assemblages were recognized by means of paleontologic, paleoecologic, and sedimentologic methods.

The paleontologic criteria for the recognition of asynchronous allochthonous complexes include the different ages of fossils composing the assemblage and dissimilarity of the taxonomic composition from that of the autochthonous assemblage. Synchronous assemblages have autochthonous and allochthonous components of the same age, but with different taxonomic compositions.

The paleoecologic criteria for the distinguishing of allochthonous complexes include the discrepancy between the habitats of organisms and their burial environments, mixture of ecologically incompatible assemblages, spatial alignment and sorting of fossils, occurrence of one or several age generations, and poor preservation of remains.

In order to avoid an erroneous interpretation, we should take into account that the preservation of fossils is evaluated by considering the shell damage, corrosion, fragmentation, articulation of a skeleton, traces of transportation, and roundness of originally not rounded fossils. The poor preservation may result from both the transportation of remains with their repeated treatment and the predator activity, bioturbation, and prolonged exposure of remains on the sea floor before their burial. The subsequent taphonomic processes, including fragmentation, breakdown of compound skeletons, limited leaching, shell dissolution, replacement by pyrite, dolomite, and anhydrite, as well as the calcite deposition in

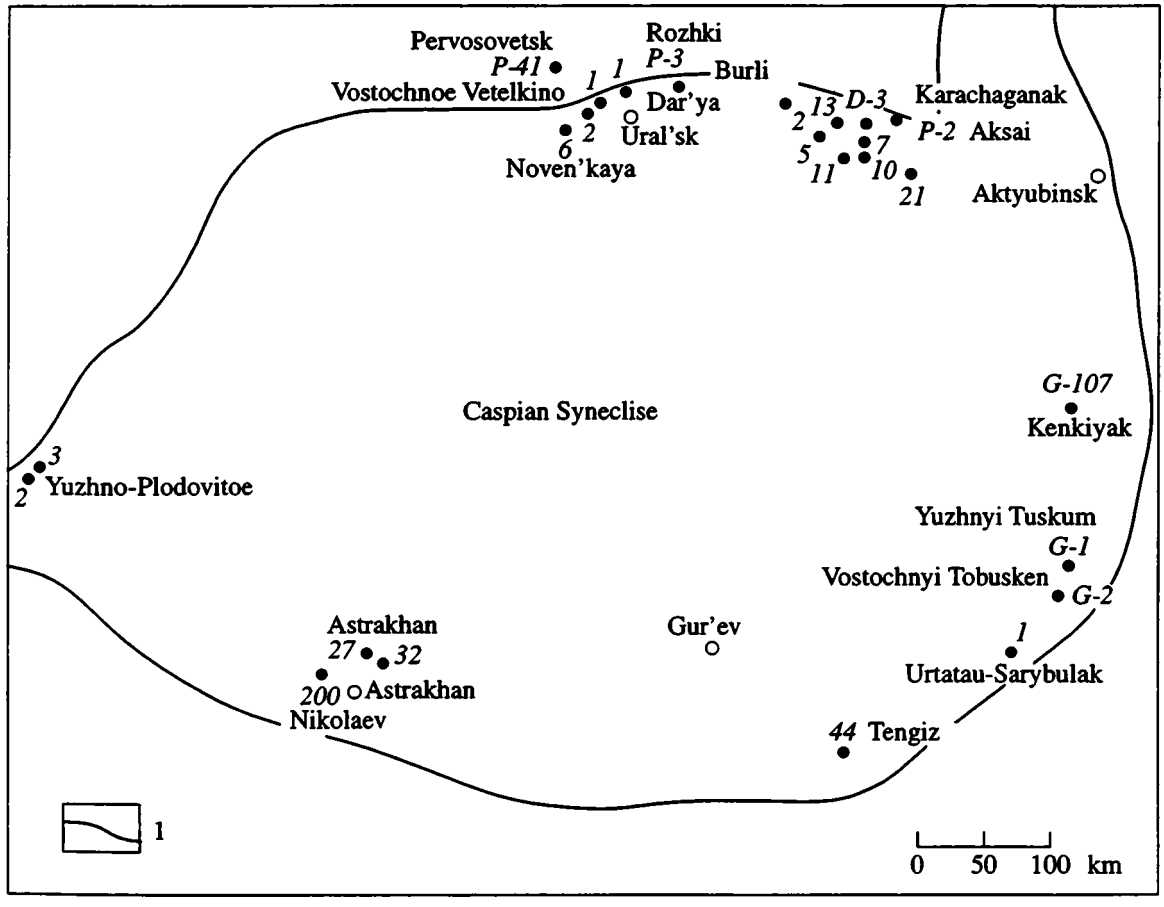

Fig. 1. Location of boreholes filled circles with allochthonous benthic assemblages in the Caspian syneclise. 1, boundary of the Caspian syneclise.

Comparison of distinguishing criteria for asynchronous and synchronous allochthonous benthic assemblages

Criteria	Redeposited benthic fossils	
	synchronous (S)	asynchronous (A)
Fossils of different age	–	+
Age difference with respect to autochthonous assemblage	–	+
Difference of taxonomic composition as compared to autochthonous assemblage	+	+
Fossils of a single age generation	+	–
Fossils of different age generations	+	+
Discrepancy between facies of habitat and burial of fossils	+	+
Mixture of ecologically incompatible assemblages	+	+
Damaged fossils, traces of transportation, roundness	+	+
Uncoupled shell valves	+	+
Spatial orientation of fossils	x	x
Sorting of fossils according to size	x	x
Different composition of enclosing and shell-filling sediments	+	+
Intraclasts	+	x
Extraclasts	–	+
Synchronous ooids of highly energetic waters and allochthonous origin	+	–
Ooids from older deposits	–	+
Synchronous allochthonous oncoids	x	–
Erosional structures	x	+
Slump and brecciated structures	x	x
Small-scale crisscross-bedding, graded bedding, and tile structure	x	–
Plucking traces, and excavation marks	x	–
Poorly sorted matrix	x	–

Note: (+) present; (x) present in some facies; (–) not present.

shell hollows, also affected the fossil preservation. The roundness of fossils originally not rounded, as well as of bioclasts (Flugel, 1982), depends on the fossil abundance and size. At a high frequency of occurrence, not all fossils and bioclasts are rounded even in energetic conditions. Small fossils are mainly rounder than large ones. The degree of preservation of benthic fossils tends to decrease progressively with distance from the initial ecologic niche. At the same time, fossils transported for substantial distances during short storms may have a perfect preservation.

The sedimentologic criteria for distinguishing allochthonous assemblages include (1) the occurrence of redeposited rock fragments, namely, intraclacts and extraclasts, as well as ooids, and allochthonous oncoids in enclosing rocks, (2) the different composition of enclosing sediments and those filling shells, (3) the poor sorting of matrix material, (4) the presence of erosional, slump, and brecciated structures, as well as cross-bedding, and (5) indicators of storm activity, namely, plucking traces and excavation marks. Sedimentologic redeposition indicators may be smoothed out by postsedimentation processes. We used the classification of R. Dunham (1962) for carbonate rocks.

Diagnostics of allochthonous assemblages should be based on the critical evaluation of all available evidence. Different criteria have major importance for various localities. The indicators mentioned above, in whole or in part, are most distinct for the assemblages transported far from their habitats. In parautochthonous complexes, redeposition marks are poorly expressed.

It should be emphasized that a great deal of the criteria for distinguishing allochthonous assemblages mentioned above are identical with those established for transported particles (Flugel, 1982), namely: (a) angular or rounded intraclasts and extraclasts, (b) rounded fragments of initially not rounded fossils, (c) poorly sorted matrix, (d) mixture of carbonate and siliceous particles of the same size, (e) assemblages of ecologically incompatible organisms, (f) presence of ooids (this criterion is used with caution, as ooids are formed not only by wave activity, but in quiet water as well), (g) wave-resistant colonial organisms *in situ*, (h) characteristic sedimentation tex-

tures, namely, the small-scale crisscross-bedding, graded bedding, and tile textures in beach sediments.

A correlation of criteria for the distinguished redeposited benthic assemblages is cited in the table. Attention is drawn to the similarity of the major sedimentologic and paleoecologic criteria of the two types of assemblages. They differ mainly by the same age of fossils in synchronous complexes and by the different age of them in asynchronous assemblages. Other important differences are presented by the following sedimentologic criteria: (1) Asynchronous allochthonous complexes are characterized by the presence of older rock fragments or extraclasts, whereas synchronous ones have partly consolidated intraclasts of the same age or allochems. The exception is the submarine slope facies, where asynchronous assemblages are accompanied not only by extraclasts, but by intraclasts as well. (2) Asynchronous assemblages may occur together with older ooids, whereas synchronous ones occur with ooids of the same age from highly energetic waters, as well as with allochthonous ooids and oncoids.

The lateral distribution of allochthonous assemblages can be compared with seed scattering (Bigon *et al.*, 1989), which is governed by the inverse square and inverse cube laws.

According to the inverse square law, skeletal remains, like seeds, are scattered unevenly, and their localities are isolated and remote from each other. The abundance of allochthonous fossils is the greatest at the smallest distance from the habitat, and sharply decreases outward from it. According to the inverse cube law, allochthonous assemblages were distributed in clearly defined beds, at the base of transgressive cycles.

SYNCHRONOUS ALLOCHTHONOUS ASSEMBLAGES

The formation of synchronous assemblages is associated with cyclic and eventual processes.

Eventual or turbulent processes produced the lateral transportation of benthic organisms from shallow-water facies to relatively deep-water ones.

Allochthonous benthic assemblages from Well 21 (interval from 5001 to 4999 m) in the Karachaganak area are examples of such horizontal transportation. Here, lower Bashkirian relatively deep-water facies (Klenina and Ovnatanova, 1986) of mostly dark-gray limestones, siltstones, and shales, yield conodonts of the *Declinognathodus* background biofacies. At the same time, within this relatively deep-water facies, considered to be unfavorable for zoobenthos, the microbedded limestone layer without bioturbation contains the foraminifer assemblage with *Tuberitina maljavkini* Mikh., *Palaeonubecularia uniserialis* Reitl., *Archaediscus pseudomoelleri* Reitl., *Asteroarchaediscus baschkiricus* (Krest. et Theod.), *A. rugosus* (Raus.), *A.* cf. *ovoides* (Raus.), *Neoarchaediscus* ex gr. *postrugosus* (Reitl.), *Eolasiodiscus* ex gr. *donbassicus* Reitl., *Endothyra bowmani* Phillips, *Mediocris breviscula* (Gan.), and *Millerella* sp., whose distribution in the studied area is limited to shoals. The transportation and secondary burial of this foraminifer assemblage can be explained by the activity of a turbidity current or strong water flows during the storm peak, when a combined interaction of wind and currents with storm and orbital waves occurred, and the wave base was substantially lower than usual. The evidence for the action of storm waves is, in this case, the presence of a smooth surface with plucking traces and excavation marks and the characteristic indicators of short-term highly energetic processes at the base of the foraminifer bed. The underestimated lateral transportation of the foraminifer assemblage can result in an erroneous paleoecologic conclusion that the above-mentioned complex is a paleobiocoenosis.

The allochthonous benthic assemblages generated by cyclic processes were recorded in the facies of lagoons, wave zone shoals (shoal carbonate sands), organogenic buildups, open shelf, submarine slope, submarine fans, as well as in abyssal facies.

In the studied region, synchronous allochthonous assemblages occur in a wide stratigraphic range from the Lower Devonian to the Lower Permian.

According to the relationship between components, including allochthonous skeletal remains, allochthonous assemblages are divided into the following six groups occuring in: (1) intraclasts, (2) ooids and oncoids, (3) ooids and intraclasts, (4) intraclasts and matrix, (5) ooids, intraclasts, and matrix, and (6) matrix.

Allochthonous Assemblages in Intraclasts

Allochthonous complexes of benthic fossils included in intraclasts occur in the Lower Devonian, Serpukhovian, and Lower Bashkirian sediments from the northern flank zone, represented by lagoon facies and sorted carbonate sands from shoals, and the open shelf facies.

Lower Devonian (Emsian) sediments from Well D-2 in the Karachaganak area, recovered in the interval from 6417 to 6412 m, are represented by open shelf facies composed of microcrystalline and rarer finely crystalline limestones containing few intraclasts of pelitomorphic limestones. They yielded two foraminifer assemblages—the autochthonous in the main body, and the allochthonous in fragments of pelitomorphic limestones. The autochthonous assemblage dominates, and includes foraminifers *Parathurammina* sp., *Baituganella* sp., and *Cribrosphaeroides irregularis* Pron. The allochthonous component is represented by the single species *Parathurammina* aff. *kolongensis* Pron.

In the Serpukhovian sediments (Protva Horizon) recovered by Well 6 in the Noven'kaya area, mixed

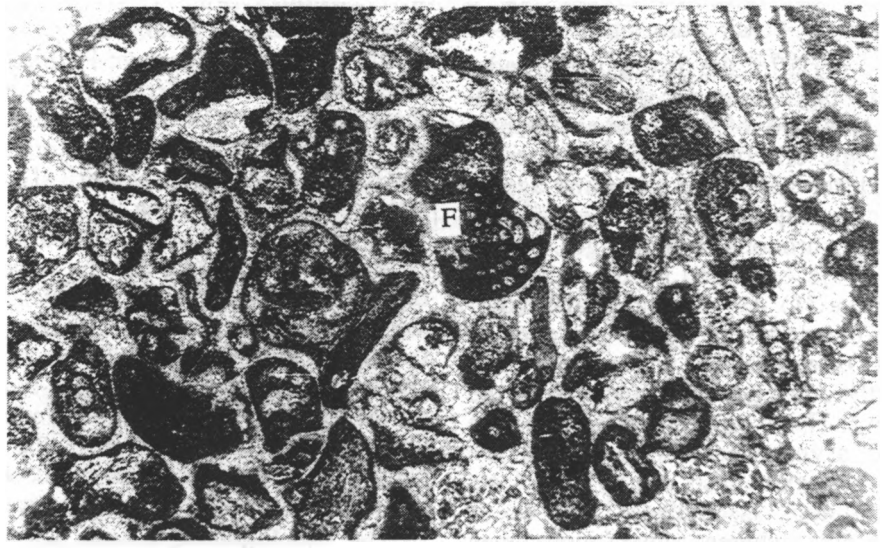

Fig. 2. Bioclastic limestone (wackestone). Allochthonous foraminifer *Eostaffella ikensis* Viss. (F; ×26) in the intraclast. Well 6 in Noven'kaya area, thin section 9149, interval 4732–4725 m. Lower Carboniferous, Serpukhovian Stage, upper Serpukhovian Substage, Protva Horizon.

assemblages were recorded in the lagoon facies at two levels. In the lower interval from 4779 to 4774 m, bioclastic limestones (packstones) and grumous limestones (pelleted wackestones) contain the allochthonous foraminifer *Eostaffella* cf. *ikensis* Viss., whose shells, initially buried in the calcareous ooze, are now components of intraclasts included in the matrix of the rock. In the upper interval from 4732 to 4725 m, bioclastic limestones, or wackestones, yielded intraclasts of bioclastic limestones (packstones–wackestones) with the redeposited foraminifers *Endothyra* sp. and *Eostaffella ikensis* Viss. (Fig. 2), as well as with fragments of calcareous algae *Calcifolium okense* Shw. et Bir. and *Koninckopora* cf. *inflata* (Koninck.).

The allochthonous assemblage, also incorporated into the mixed complex, was also found upwards in the section of this well (interval from 4629 to 4619 m, thin section 9128) namely, in lower Bashkirian (Severnaya Kel'tma) clastic limestones with intraclasts of algal, *Donezella* limestones, as well as of bioclastic, crinoidal, and oolitic limestones. The allochthonous components in intraclasts are represented by the foraminifers *Pseudoglomospira gordialis* (Jon. et Park.), *Neoarchaediscus subbaschkiricus* (Reitl.), *Eostaffella pseudostruvei* Raus. et Bel., *Plectostaffella* cf. *bogdanovskensis* Reitl., and *P.* sp.

Allochthonous Assemblages in Ooids and Oncoids

Allochthonous assemblages associated with ooid and oncoid cores are characteristic of the facies of sorted carbonate shallow-water sands of the lower Tournaisian, lower Bashkirian, and Asselian in the northern flank zone.

In the lower Tournaisian sediments from Well 7 in the Karachaganak area (the interval from 5094 to 5089 m), oolitic and oncoid limestones of the sorted carbonate shallow-water sands yielded redeposited foraminifers, namely, endothyrids, parathuramminids, and tournaiellids, together with calcareous algae fragments. Shells of free-lying foraminifers and stalks of sessile calcareous algae, which inhabited the quiet lagoon, were transported to shoals and bars by tidal or directional currents. After repeated transportation under the conditions of a high and moderately high hydrodynamic regime, the skeletal remains became cores of ooids and oncoids.

The most taxonomically and quantitatively rich allochthonous benthic components of mixed assemblages and the proper allochthonous complexes were recorded in lower Bashkirian sediments recovered by Well 6 in the Noven'kaya area (Fig. 3) at eight levels of the Krasnaya Polyana and five levels of the Severnaya Kel'tma horizons. The lower Bashkirian deposits, including the indicated horizons, are mainly composed of oolitic limestones with few interlayers of oncoid and clastic limestones. Allochthonous remains of foraminifers, gastropods, ostracodes, crinoids, and calcareous algae occur primarily in ooid or oncoid cores.

In the Krasnaya Polyana sediments, the allochthonous assemblage within the mixed complex was recorded in oolitic limestones (interval from 4670 to 4663 m, thin section 9147), which contain few lithoclasts of bioclastic limestones. Here, foraminifers *Baituganella vulgaris* Lip. (?), *Tolypammina* sp., *Paracalligella* sp., *Eostaffella amabilis* Grozd., and calcare-

ous algae occur in ooid cores. *Tolypammina* sp. is a form in common with the autochthonous assemblage.

Another allochthonous assemblage (thin section 9146), also included in the mixed complex, was found 1.5 m upward the section in oolitic limestones with a characteristic cross-bedding (Fig. 4). Ooid cores contain the foraminifers *Archaesphaera minima* Sal., *Trepeilopsis* cf. *grandis* Brazhn., *Neoarchaediscus gregorii* (Dain.), *Globivalvulina* ex gr. *mosquensis* Reitl., and calcareous algae *Girvanella* sp., and *Fasciella* sp.

The allochthonous assemblage (thin section 9145) within the mixed complex in the studied interval is associated with oolitic limestones yielding in ooid cores the foraminifers *Mediocris mediocris* (Viss.), *M. breviscula* (Gan.), and calcareous algae *Praedonezella* aff. *cespeformis* Kulik. (Fig. 5).

Allochthonous components of the mixed complex were found in the interval from 4656 to 4649.5 m, thin section 9143, in algal *Donezella* limestones with few abraded ooids. Calcareous algae *Donezella lutugini* Masl. occured in ooid cores.

Allochthonous foraminifers were also recorded 0.3 m upward the section in oolitic limestones with intraclasts of silty, bioclastic, algal–oolitic, and algal–foraminiferal limestones (thin section 9142). The foraminifer shells of *Eostaffella* cf. *postmosquensis* Kir. represent some ooid cores.

The proper allochthonous assemblage was recorded in the interval from 4649.5 to 4644 m, thin section 9141, in oolitic limestones containing lithoclasts of bioclastic and micrograined limestones. The assemblage includes the foraminifers Textulariidae gen. et sp. indet., *Plectostaffella* sp., and *Plectomillerella* cf. *subacuta* Brazhn. et Vdov., which form ooid cores.

The more diverse allochthonous assemblage, a component of the mixed complex (thin section 9138), was found 3.5 m upwards the section in oolitic limestones. Ooid cores contain the foraminifers *Tuberitina maljavkini* Mikh., *Tolypammina fortis* Reitl., *Haplophragmina beschevensis* Brazhn., Textulariidae gen. et sp. indet., *Eostaffella* cf. *postmosquensis* Kir., and *E.* cf. *pseudostruvei* Raus. et Bel., as well as calcareous algae *Fasciella kizilia* R. Ivan., *F. ivanovae* Saltovskaja, *Donezella lutugini* Masl., and *Dvinella comata* Khvorova. Among them, *Tuberitina maljavkini* Mikh. and *Donezella lutugini* Masl. are in common with the autochthonous assemblage.

One more allochthonous assemblage, also a part of the mixed complex, was recorded in oolitic limestones 1 m upward the section, thin section 9137. It is the most abundant and rich in taxonomic composition. The following foraminifers and calcareous algae were identified in ooid cores: *Tuberitina collosa* Reitl., *Ammovertella delicata* Reitl., *A. vaga* Reitl., *A.* cf. *lata* Reitl., *Pseudoglomospira gordialis* (Jon. et Park.), *Asteroarchaediscus baschkiricus* (Krest. et Theod.), *A. ovoides* (Raus.), *Neoarchaediscus subbaschkiricus* (Reitl.), *N. gregorii* (Dain), *N. postrugosus* (Reitl.), *Eostaffella*

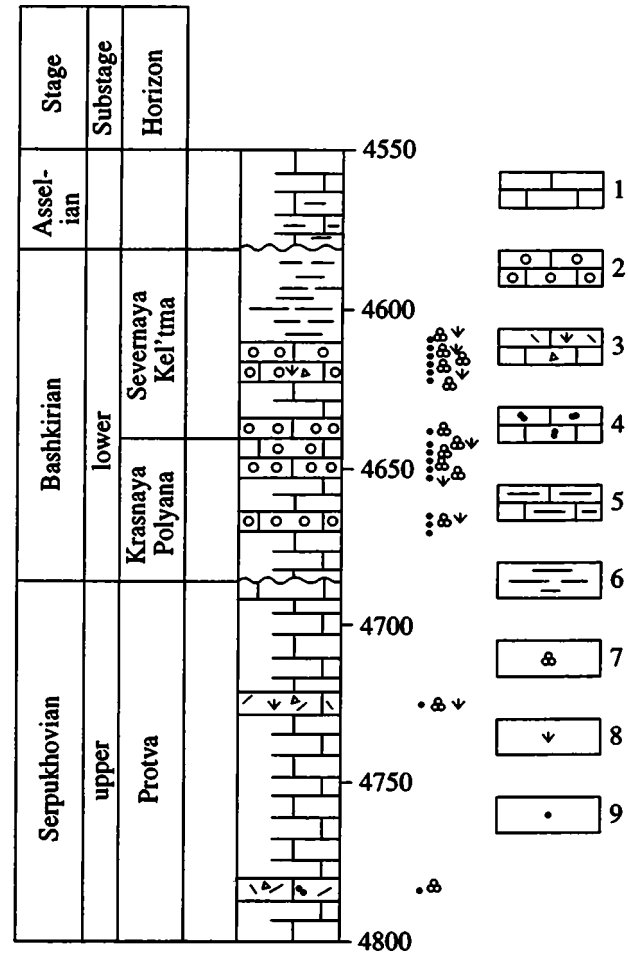

Fig. 3. Section of Carboniferous sediments of Well 6 in the Noven'kaya area: 1, limestones; 2, oolitic limestones; 3, bioclastic limestones; 4, grumous limestones; 5, clayey limestones; 6, argillites, 7, foraminifers; 8, algae; 9, location of allochthonous assemblages.

postmosquensis Kir., *Plectostaffella* sp., and *Fasciella kizilia* R. Ivan. Among them, foraminifers *Ammovertella delicata* Reitl., *Asteroarchaediscus baschkiricus* (Krest. et Theod.), and *Eostaffella postmosquensis* Kir. occur in the autochthonous assemblage as well.

Allochthonous foraminifer assemblages from the Krasnaya Polyana deposits recovered by Well 6 in the Noven'kaya area yielded 38 species, four of which (namely, *Neoarchaediscus gregorii* (Dain), Textulariidae gen. et sp. indet., *Eostaffella* cf. *postmosquensis* Kir., and *Plectostaffella* sp.) occur at two levels. The other forms are present at a one certain level.

Among the Krasnaya Polyana allochthonous algal assemblages, only two have the common species *Fasciella kizilia* R. Ivan. and *Donezella lutugini* Masl.

The overlying deposits of the Severnaya Kel'tma Horizon contain an impoverished allochthonous assemblage in the interval from 4644 to 4636 m, thin section 9135. Ooid cores of oolitic limestones yielded

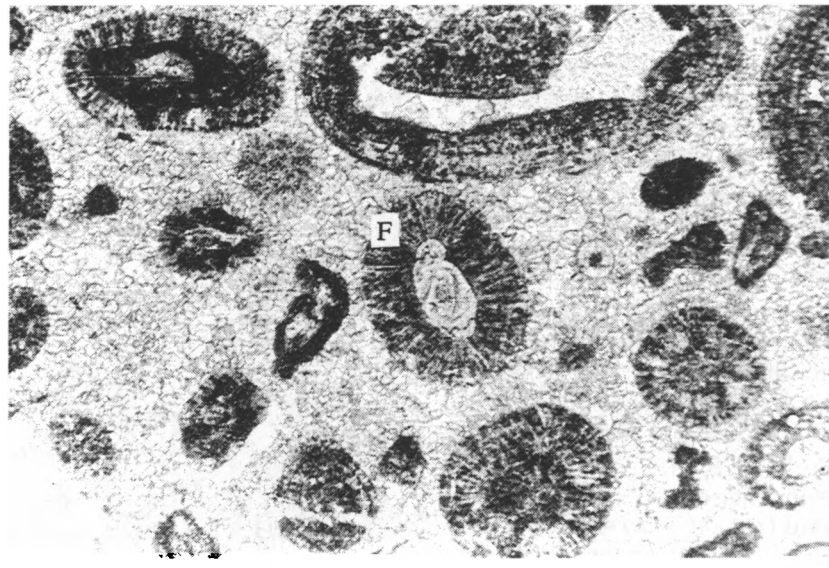

Fig. 4. Oolitic limestone. Allochthonous foraminifer *Neoarchaediscus gregorii* (Dain) (F; ×57) in the ooid core. Well 6, Noven'kaya area, thin section 9146, interval 4670–4663 m. Middle Carboniferous, Bashkirian Stage, lower Bashkirian Substage, Krasnaya Polyana Horizon.

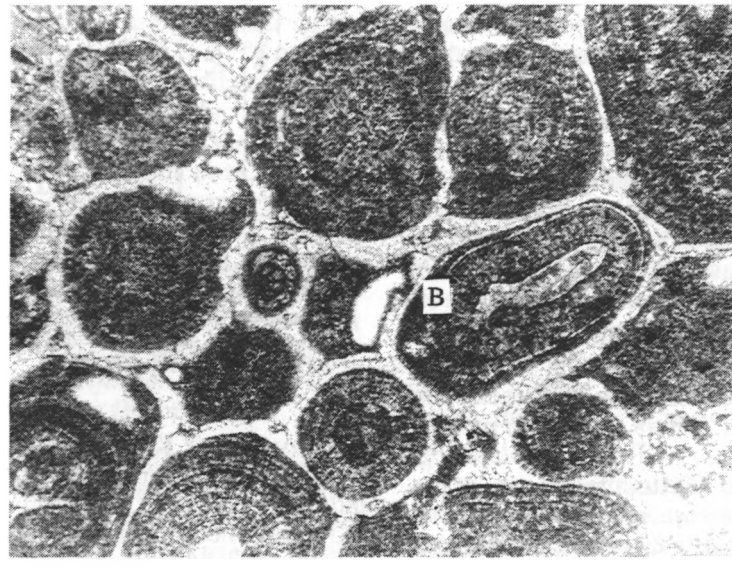

Fig. 5. Oolitic limestone. Allochthonous algae *Praedonezella* aff. *cespeformis* Kulik (A; ×46). Well 6, Noven'kaya area, thin section 9145, interval 4670–4663 m. Middle Carboniferous, Bashkirian Stage, lower Bashkirian Substage, Krasnaya Polyana Horizon.

the poorly preserved foraminifers *Eostaffella* sp. and *Palaeotextularia* sp.

The next properly allochthonous assemblage was recorded 3.7 m upward the section (thin section 9132), where the calcareous algae *Koninckopora* sp. occurred in oncoid cores. Relict oncoid limestones represent the enclosing sediments.

It should be marked that this allochthonous assemblage is somewhat conventionally considered as synchronous, as it contains the allochthonous algae *Koninckopora* distributed in underlying Lower Carboniferous sediments. It is not inconceivable that this assemblage is asynchronous with the enclosing deposits, but the poor preservation of algae prevents the formulation of an unambiguous conclusion, because the flourishing of *Koninckopora* representatives occurred in the late Visean. They were practically extinct by the beginning of the Serpukhovian, though the latest forms occurred outside the studied region even in the Bashkirian.

Allochthonous components are associated with oolitic limestones in the interval 4619–4611 m, thin

section 9126. Foraminifers *Pseudoglomospira gordialis* (Jon. et Park.), *Pseudostaffella* cf. *antiqua* (Dutk.), and *P.* sp. were found in ooid cores.

The more diverse allochthonous assemblage within the mixed complex occurs 1.5 m upward the section (thin section 9127-1) in oolitic limestones, where ooid cores yielded the foraminifers *Paracalligella* sp., *Pseudoglomospira gordialis* (Jon. et Park.), *Palaeonubecularia uniserialis* Reitl., *Neoarchaediscus subbaschkiricus* (Reitl.), *Eostaffella postmosquensis* Kir., *Plectostaffella varvariensis* (Brazhn. et Pot.), *P. bogdanovkensis* Reitl., *Pseudostaffella* cf. *antiqua* (Dutk.), *P. antiqua grandis* (Schlyk.), *P. compressa* Saf., and calcareous algae *Koninckopora* sp.

An allochthonous assemblage of somewhat different composition was recorded 0.1 m upward (thin section 9127-2). Oolitic limestones contained intraclasts of the same rocks, and their cores yielded the foraminifers *Tolypammina* ex gr. *fortis* Reitl., *Asteroarchaediscus baschkiricus* (Krest. et Theod.), *Neoarchaediscus gregorii* (Dain), *N.* cf. *postrugosus* (Reitl.), *Eostaffella postmosquensis* Kir., *E. pseudostruvei chomatifera* Kir., *E. pseudostruvei angusta* Kir., *Pseudostaffella antiqua* (Dutk.), and calcareous algae *Donezella* cf. *lunaensis* Racz.

Comparative studies of allochthonous foraminifer assemblages from the Krasnaya Polyana and Severnaya Kel'tma horizons revealed the following eight common species: *Paracaligella* sp., *Pseudoglomospira gordialis* (Jon. et Park.), *Asteroarchaediscus baschkiricus* (Krest. et Theod.), *Neoarchaediscus subbaschkiricus* (Reitl.), *N. gregorii* (Dain), *N. postrugosus* (Reitl.), *Eostaffella postmosquensis* Kir., and *Plectostaffella* sp.

Autochthonous foraminifer assemblages from the Krasnaya Polyana and Severnaya Kel'tma horizons yielded six common species, namely, *Palaeonubecularia uniserialis* Reitl., *Bradyina* ex gr. *cribrostomata* Raus. et Reitl., *Eostaffella parastruvei* Raus., *E.* cf. *postmosquensis* Kir., *Semistaffella variabilis* (Reitl.), and *S.* cf. *primitiva* (Reitl.), among which only *Asteroarchaediscus baschkiricus* (Krest. et Theod.) and *Eostaffella postmosquensis* Kir. occur in the allochthonous assemblages of studied deposits. The active zone of sand transportation and accumulation in the oolitic shoal of the Lily Bank located parallel to the northeastern margin of the Small Bahama Bank (Hine, 1977) is, evidently, analogous to the oolitic shoal of the Noven'kaya area.

In the Asselian sediments recovered by Well 5 in the Karachaganak area in the interval from 5036 to 5028 m, oolitic limestones with normally packed grains contain the allochthonous assemblage associated with ooid cores and represented by foraminifer shells and their poorly preserved fragments.

Allochthonous Assemblages in Ooids and Intraclasts

The allochthonous benthic assemblages included in ooids and intraclasts were recorded in the Bashkirian sediments from the northern and southwestern flank zones.

In the northern flank zone, the allochthonous component of a mixed assemblage was discovered in the Severnaya Kel'tma sediments recovered by Well 6 in the Noven'kaya area. It occurs in the facies of sorted carbonate shallow-water sands in the interval from 4629 to 4619 m (thin section 9133) and in oolitic limestones with intraclasts of bioclastic and grumous limestones. The allochthonous foraminifers *Eostaffella* cf. *pseudostruvei angusta* Kir. and *Archaediscus* sp. were found in ooid cores and intraclasts, respectively.

In the southwestern marginal zone, Bashkirian sediments represented by the facies of wave zone shoals were recovered by Well 3 in the Yuzhno-Plodovitoe area. Two allochthonous assemblages—components of the mixed complexes—were distinguished there.

The lower assemblage was recorded in the interval from 4825 to 4812 m, where light-gray relict oolitic limestones bear fossils within intraclasts of 0.3–5 mm in size, and ooids of 0.05–0.4 mm in size. Rounded and subrounded intraclasts are represented here by algal, bioclastic, and oolitic limestones, and contain the rare foraminifers *Pseudoglomospira* sp. and *Eostaffella* cf. *pseudostruvei* Raus. et Bel. together with the calcareous algae *Fasciella* cf. *kizilia* R. Ivanova and *F.* cf. *ivanovae* Saltovskaja. Cores of ellipsoidal ooids, often micritic, yielded foraminifers, crinoids, and algae, whose identification even at the generic level is impossible. The autochthonous foraminifer assemblage differs from the allochthonous one, and includes *Neoarchaediscus postrugosus* (Reitl.), *Bradyina cribrostomata* Raus. et Reitl., *Cribrostomum* sp., *Globivalvulina* sp., *Eostaffellina paraprotvae* (Raus.), *Eostaffella postmosquensis* Kir., and *Plectostaffella varvariensis* (Brazhn. et Pot.). The presence of the autochthonous algae *Fasciella ivanovae* Saltovskaja suggests that the formation of the assemblage occurred in the less active shoal area or in another stable setting.

The upper assemblage, recorded 17 m upward the section in the interval from 4797 to 4793 m, is associated with the similar oolitic and relict oolitic limestones, but differs from the lower assemblage. Limestones include intraclasts of algal, bioclastic, and algal–foraminifer limestones with the foraminifers *Eostaffella* cf. *postmosquensis* Kir. Cores of ellipsoidal, somewhat micritic; ooids with tangential texture contain foraminifer shells of different morphology, namely, *Eostaffella* sp. and *Globivalvulina* sp., and include fragments of crinoid and calcareous algae stalks. Reliably identified allochthonous foraminifers are similar to those in the autochthonous assemblage.

Allochthonous Assemblages in Intraclasts and Matrix

These assemblages were found in upper Visean and Serpukhovian sediments from the northern flank zone and in Artinskian deposits from the southwestern marginal area.

Upper Visean sediments recovered by Well 2 in the Burli area, interval from 4284 to 4280 m, are represented by facies of the wave zone shoals. Respective clastic limestones with rare ooliths (grainstones) yielded a mixed foraminifer and calcareous algae assemblage in intraclasts and matrix. The diverse autochthonous foraminifer fauna contains late Visean *Archaediscus agapovensis* Iv., *Asteroarchaediscus bashkiricus* (Krest. et Theod.), *Omphalotis omphalotis* (Raus. et Reitl.), *O. samarica* (Raus.), *Globoendothyra globulus* (Eichw.), *Endothyranopsis compressa* (Raus. et Reitl.), *Eostaffella ikensis* Viss., *E. proikensis* Raus., *E. parastruvei* Raus., and *Parastaffella propinqua* Viss. Allochthonous foraminifers from intraclasts of bioclastic limestones (grainstones and packstones) are poor, and include *Forshiella prisca* Mikh., which is absent in the autochthonous assemblage, and *Globoendothyra* sp. (Fig. 6). The autochthonous algae flora is represented by *Calcifolium okense* Schv. et Bir., *C. punctatum* Masl., *Koninckopora delicata* Nakai et Kato, and *Fasciella* sp. Its allochthonous component includes rounded or slightly rounded, often with micritic edging stalk fragments of late Visean representatives of the *Koninckopora* genus, namely, *K. delicata* Nakai et Kato, in common with the autochthonous assemblage, as well as *K. minuta* Weyer and *K.* sp. Foraminifers *Omphalotis omphalotis* (Raus. et Reitl.) and *Mediocris breviscula* (Gan.) were recorded inside one of *Koninckopora* aggregates (Fig. 7).

Serpukhovian sediments recovered by Well 13 in the Karachaganak mineral deposit area are represented by the submarine slope facies composed of bio- and lithoclastic breccia, as well as of bioclastic and microbrecciated limestones. They contain fragments of reef limestones of the same age with rich benthic fauna, algae flora, and bioclasts in the intervals of 5089–5083, 5077–5072, and 5072–5068 m. Allochthonous material is chiefly represented by skeletal fragments of corals and calcareous algae, evidently damaged by bioerosion and broken by storms. Some bioclasts were formed during the posthumous desintegration of skeletons of framework constructors and reefophils transported down the slope. The slight roundness of fragments indicates their local origin and excludes the long transportation. The distance, along which the reef material was moved, did not exceed several kilometers, and this agrees with the data on modern reefs. For instance, in the Princess Charlotte Bay near the western Australian coast, the reef material occurs in remote deposits not farther than at 2 km from the reef (Frankel, 1974).

Thus, the presence of reef fauna and reefal rock fragments in the slope facies may suggest the proximity of a reef complex. Allochthonous benthic assemblages from old slope facies also permit the reconstruction of the initial structure of submarine margins of a carbonate platform or that of reef carbonates later altered by postsedimentation processes.

The youngest synchronous allochthonous assemblages were found in Artinskian sediments recovered by Well 2 from the Yuzhno-Plodovitoe area and represented by the submarine slope facies.

The first assemblage was recorded at the base of the Artinskian section, where the interval from 3463 to 3456 m is marked by dark-gray brecciform limestones containing angular and subangular clasts from 1–5 to 16 mm in size. The following types of clasts were distinguished: (1) organoclastic limestone with the foraminifers *Palaeonubecularia uniserialis* Reitl., *Climacammina* sp., *Pseudoendothyra* cf. *preobrajenskyi* (Dutk.), Schwagerinidae gen. et sp. indet., bryozoans, and calcareous algae *Tubiphytes* sp.; (2) bioclastic limestone with the foraminifers *Hemigordius* sp., *Schubertella* sp., Schwagerinidae gen. et sp. indet., and algae *Tubiphytes* sp.; (3) lump-grumous limestone with the foraminifer *Schubertella* (?) sp. and algae *Tubiphytes* sp.,; (4) microcrystalline silty limestone with a few fusulinids and the algae *Tubiphytes* sp.; (5) relict organogenic limestone; and (6) calcareous aleuropelite. The matrix of brecciform limestones is represented by fine-grained crystalline calcite and secondary dolomite. The matrix of the rock yielded rounded fragments of Schwagerinidae shells and *Tubiphytes* sp. stalks, which indicates their transportation before the burial.

Another assemblage occurs in the upper part of the Artinskian sediments (interval from 3414 to 3410 m) in bioclastic dolomitic limestones (grainstones), and in limestones with a prevailing matrix (flowstones). The allochthonous assemblage occurs both in the matrix and in single intraclasts. Angular and subangular fragments of grumous limestones which yielded the foraminifers *Ammodiscus* sp. Bioclasts in the matrix are represented by rare broken shells of the foraminifers *Ammodiscus* sp., *Pseudoendothyra* sp., *Ozawainella* sp., *Schubertella* sp., and *Pseudofusulina* sp., bryozoans *Rhombotrypella* and *Fenestella*, segments of crinoid stalks, as well as by the calcareous algae *Tubiphytes* cf. *obscurus* Masl. and *T.* sp.

Comparison of the two assemblages mentioned above shows that they have such taxa in common as pseudoendothyrids, schubertellids, and tubiphytes, whereas the main differences are evident from the confinement of one or another assemblage mainly to intraclasts or matrix, respectively, as well as from the larger diversity of redeposited foraminifers in the first assemblage.

Allochthonous Assemblages in Ooids, Intraclasts, and Matrix

In this case, three components of rock enclosing allochthonous assemblages were recorded in the single section of the southwestern marginal zone, in the facies of sorted shallow-water carbonate sands.

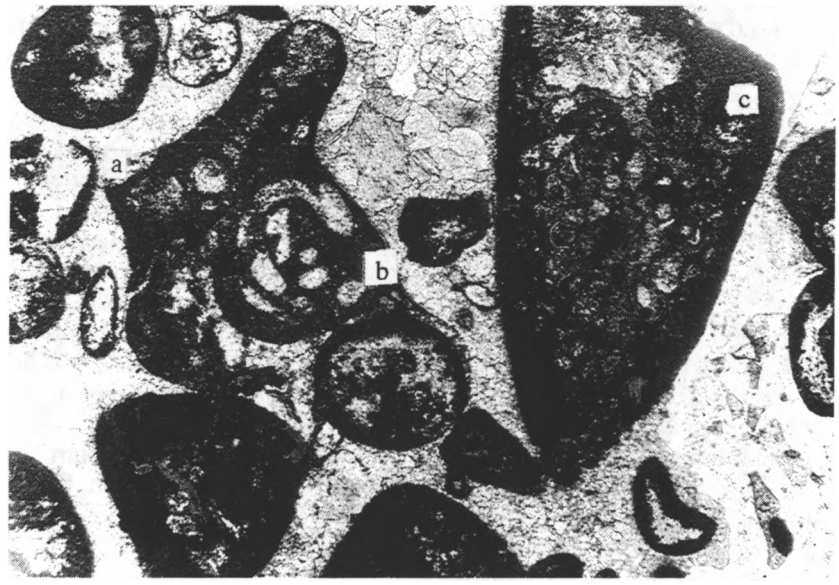

Fig. 6. Clastic limestone (grainstone). Allochthonous foraminifers *Forshiella prisca* Mikh. (a) and *Globoendothyra* sp. (b) in the intraclast, and allochthonous calcareous algae *Koninckopora* sp. (c; ×25) in the matrix. Well 2, Burli area, thin section 89487-1, interval 4284–4280 m. Lower Carboniferous, upper Visean.

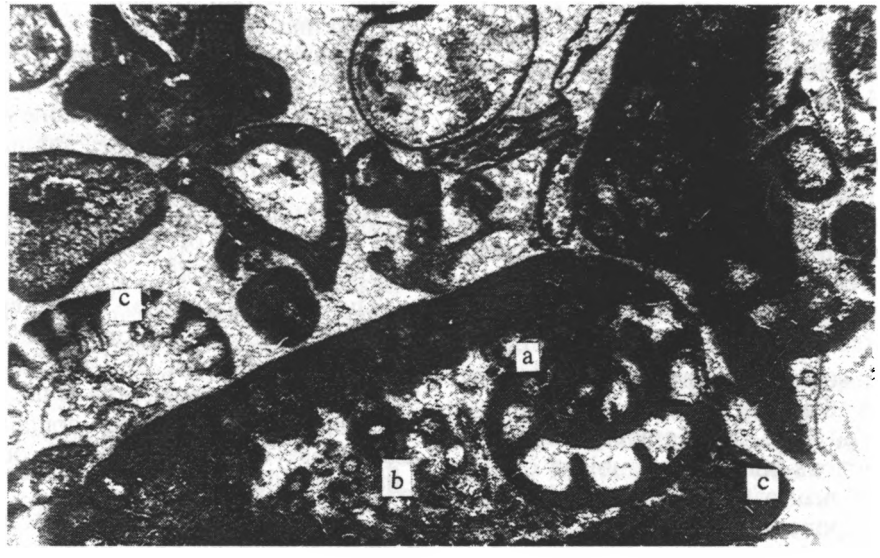

Fig. 7. Clastic limestone (grainstone). Two allochthonous fragments of calcareous algae *Koninckopora* sp. (a) in the matrix and foraminifers *Omphalotis omphalotis* (Raus. et Reitl.) (b) and *Mediocris breviscula* (Gan.) (c; ×25) inside the largest fragment. Well 2 Burli area, thin section 89487-2, interval 4284-4280 m. Lower Carboniferous, upper Visean.

In the southern margin of the Astrakhan dome, the upper part of lower Bashkirian sediments (the Kama Horizon) represented by the studied facies is recovered by Well 200 in the Nikolaev area. In the interval from 4224 to 4216 m oolitic, in places clastic limestone, or grainstone–wackestone, with enveloped grains and rare ooids, yielded the allochthonous assemblage, a component of the mixed complex. The autochthonous assemblage includes the early Bashkirian foraminifers *Eostaffella pseudostruvei chomatifera* Kir., *E. parastruvei chusovensis* Kir., *Pseudostaffella antiqua grandis* Schlyk., *P. antiqua posterior* Saf., *P. korobezkich* Raus. et Saf., and ostracodes. The allochthonous component is represented by the foraminifers *Pseudostaffella* cf. *antiqua* (Dutk.), calcareous algae *Fasciella ivanovae* Saltovskaja and *F. kizilia* R. Ivanova occuring

in ooid cores and intraclasts of bioclastic limestones, as well as by abraded, rounded and micritic, often enveloped ostracode shells, and crinoid stalk segments (Fig. 8).

Allochthonous Assemblages in the Matrix

Allochthonous benthic assemblages distributed in the matrix of the enclosing rock occur in a wide facies range of lower Visean, Serpukhovian, lower Bashkirian, and Asselian sediments.

Lower Visean calcareous siltstones from the northern flank zone, where they represent the slope facies in the section of Well 1 from the Dar'ya area (interval from 5230 to 5223 m), contain rare abraded redeposited shells of foraminifers, ostracodes, and green algae *Kamaena awirsi* Mamet et Roux.

In Serpukhovian sediments recovered by Well P-3 from the Rozhki area (interval from 3908 to 3993 m), bio- and lithoclastic limestones (wackestones), characteristic of the open shelf, yielded a redeposited assemblage of abraded and broken foraminifer shells and crinoid stalk segments, accompanied by intraclasts of bioclastic and pelitomorphic limestones (mudstones).

Serpukhovian deposits, represented by the facies of wave zone shoals, were recovered in Well 2 from the Vostochnoe Vetelkino area in the interval from 4617 to 4607 m. Clastic limestones (grainstones) include the allochthonous foraminifers *Pseudoglomospira, Palaeonubecularia, Eostaffella,* and Textulariidae gen. et sp. indet., remains of crinoids and poorly preserved calcareous algae, as well as lithoclasts of fine-lump limestones. Upward the section in the interval from 4607 to 4597 m, clastic limestones (grainstones–packstones) contain redeposited foraminifers, crinoids, and calcareous algae, together with intraclasts of algal and bioclastic limestones (wackestones and packstones).

Lower Bashkirian sediments, represented by lagoon facies of the Krasnaya Polyana Horizon in the interval 5069–5038 m of Well 5 of the Karachaganak area, yielded the allochthonous benthic assemblage of this age, which is a component of the mixed complex. The autochthonous component includes foraminifer remains, bivalve mollusks, brachiopods, ostracodes, and green algae. The allochthonous part contains broken and abraded foraminifer and gastropod shells, together with fragments of crinoid and algae stalks. The enclosing microfacies are represented by grumous limestones, or pelleted grainstones, as well as by bioclastic and organoclastic limestones (wackestones) with few intraclasts of algal limestones. Paleoecologic data indicate that the studied allochthonous assemblages were formed in lagoons with an active hydrodynamic regime.

At the Astrakhan dome in the southwestern flank zone, lower Bashkirian sediments of the reef flat subfacies were recovered in Well 32 of the Astrakhan area (interval 4117–4110 m). Here, in bioclastic limestones, the autochthonous algae flora occurs together with the poorly sorted, abraded foraminifer shells with micritic envelopes, and the rounded fragments of crinoid stalks.

In the northern flank zone, the Asselian lagoon facies was studied in the section of Well 11 of the Karachaganak area in the interval 4581–4557 m. Their allochthonous assemblages occur in limestones with an inversed structure composed of enveloped bioclasts, namely, foraminifers and calcareous algae, which were reworked under highly energetic conditions near shoals and then buried in quiet water. Autochthonous foraminifer assemblages are subdominated by textulariids and tetrataxises. Scarce gastropods and crinoids also occur.

Thus, in the studied region, synchronous allochthonous benthic assemblages were recorded in Lower Devonian, lower Tournaisian, Visean, Serpukhovian, lower Bashkirian, Asselian, Sakmarian, and Artinskian sediments.

ASYNCHRONOUS ALLOCHTHONOUS ASSEMBLAGES

Allochthonous benthic assemblages, asynchronous with respect to enclosing sediments, are usually distributed at the base of transgressive cycles and found in the lower Bashkirian (Middle Carboniferous), Asselian and Sakmarian deposits, as well as in the lower and upper parts of the Artinskian sediments (Lower Permian). Asynchronous allochthonous assemblages mark paleoenvironments of wave zone shoals, organogenic buildups, submarine slope, submarine fans, and the open shelf.

Allochthonous skeletal material usually occurs in two components of enclosing rock, namely, in the matrix and lithoclasts.

The basal Bashkirian beds represented by slope facies yielded asynchronous allochthonous assemblages in a section of Well 13 (interval 5062–5057 m) from the Karachaganak area of the northern flank zone. Microbrecciated limestones contain the allochthonous assemblage of Early Carboniferous foraminifers and calcareous algae.

The Lower Permian sediments include an asynchronous allochthonous benthic assemblage with dominating fusulinids, whose shells are easily transportable and thus occur in different lithologic rock types.

In the slope facies of lower Sakmarian deposits, the allochthonous fusulinid assemblage, part of the mixed complex, is known in the section of Well P-2 of the Aksai area (Izotova, 1985) of the northern flank zone. The autochthonous assemblage comprises early Sakmarian fusulinids, whereas Late Carboniferous and Asselian forms compose the allochthonous complex.

In the reef facies of the upper Sakmarian sediments (the Sterlitamak Horizon), asynchronous allochthonous fusulinids were recorded in the section of Well P-10 of the Karachaganak area, interval 4242–4155 m. Together with autochthonous forms, they compose a

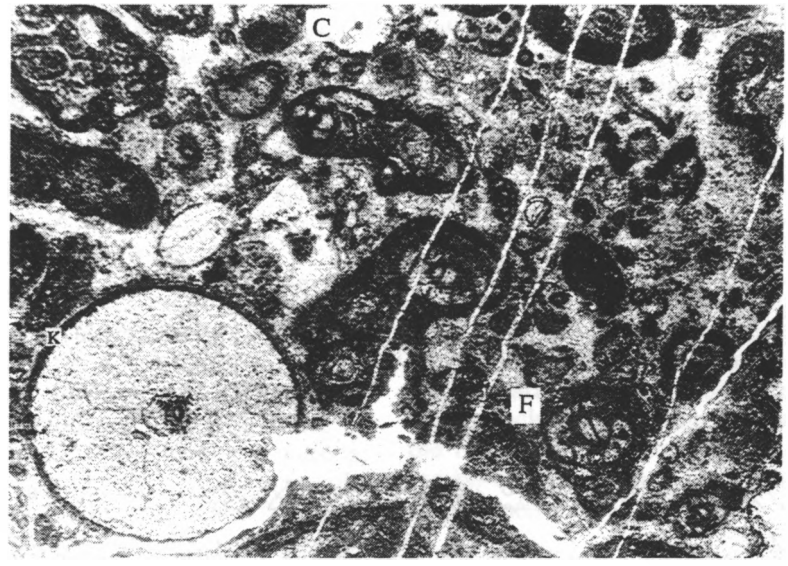

Fig. 8. Clastic limestone with enveloped grains and ooids (grainstone-wackestone). Allochthonous foraminifer *Pseudostaffella* cf. *antiqua* (Dutk.) (F) in the ooid core and allochthonous partly destroyed crinoid (C) segments with micritic edging (×26). Well 200 Nikolaev area, thin section 3-3, interval 4224–4216 m. Middle Carboniferous, Kama Horizon.

mixed assemblage. The enclosing sediments are represented by fusulinid and bryozoan limestones overlying the eroded surface of bioclastic limestones with fossils of the middle *Schwagerina moelleri–Pseudofusulina fecunda* Zone of the Asselian. The autochthonous Sterlitamak assemblage is characterized by fusulinids of the *Pseudofusulina urdalensis* Zone; it includes, along with the index species, the following forms: *Pseudofusulina urdalensis abnormis* Raus., *P. plicatissima* Raus., *P. plicatissima irregularis* Raus., *P. callosa* Raus., *P. callosa proconcavutas* Raus., and *P. ovoides* Kir. The allochthonous component includes broken fusulinid shells of the Asselian-Sakmarian age identified as *Rugosofusulina* sp., *Dutkevitchia* sp., *Daixina* sp., *Pseudofusulina* ex gr. *sphaerica* Bel., *P.* ex gr. *uralica* (Krot.), *P.* cf. *sulcata* Kozh., *P.* ex gr. *moelleri* (Schellw.), and *Schwagerina* sp. (Izotova, 1985).

Coarse clastic sediments of presumably Sakmarian age, which we believe to have been formed in submarine fans, were recovered in Well G-1 of the Yuzhnyi Tuskum area, the eastern flank zone. Redeposited fusulinid shells of different age, among which Asselian forms are the youngest, were recorded in the interval from 3456 to 3152 m (Izotova, 1985). Sakmarian deposits of the open shelf facies were recovered by Well G-107 in the Kenkiyak area. In the interval from 4240 to 4232 m, conglomerate-like sandstones contain redeposited shells of Gzhelian and late Asselian fusulinids. Artinskian sediments from the section of Well G-2 (Vostochnyi Tobusken, the interval 3263–3235 m) yielded deformed and rounded shells of Asselian and early Sakmarian fusulinids. Redeposited Carboniferous forms were identified at the Tengiz mineral deposit (Zolotukhina *et al.*, 1989) and in the Artinskian conglomerates recovered by Well 4 in the interval from 3983 to 3987 m.

FACIES CONFINEMENT OF ALLOCHTHONOUS ASSEMBLAGES

Based on the foregoing, redeposited assemblages were closely related to the sedimentation environments and characteristic of certain microfacies (Fig. 9).

Allochthonous assemblages are associated with the lagoon facies of Serpukhovian, lower Bashkirian, and Asselian sediments, where they occur in bioclastic limestones (packstones and wackestones), grumous limestones (pelleted grainstones), organoclastic limestones (wackestones), and those with inversed structures and composed of enveloped intraclasts, in the micrograined matrix.

In shoals of the wave zone—the area of transportation and accumulation of sorted shallow-water carbonate sands—allochthonous assemblages were formed in the early Tournaisian, late Visean, Serpukhovian, Bashkirian, and Asselian. Redeposited assemblages were recorded in the following microfacies: clastic and bioclastic limestones with enveloped grains (grainstones and grainstone-wackestones), coquinites, oncoid bioclastic limestones, residual breccias, and oolitic limestones corresponding to the standard microfacies of sorted shallow-water carbonate sands (Wilson, 1980). Autochthonous complexes from this facies are characterized by impoverished benthic assemblages, which inhabited the mobile bottom. The allochthonous component is mainly represented by fossil remains of organisms from adjacent facial belts. Allochthonous shells are significantly reworked, often rounded, abraded, corroded by borers, micritized, and enveloped by the algal mass. Skeletons of sessile and slightly vagile benthos were "elutriated" in shoals and became ooid and oncoid cores. A great deal of localities show con-

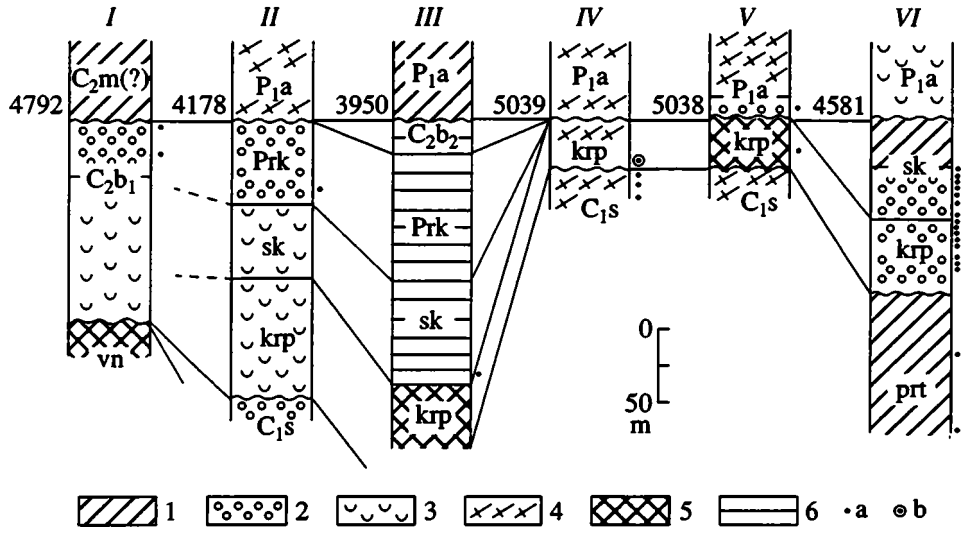

Fig. 9. Correlation of Carboniferous and Lower Permian sediments enclosing allochthonous benthic assemblages. Correlation is based on autochthonous foraminifer and algae assemblages: (*I*) Well 3, Yuzhno-Plodovitoe area; (*II*) Well 200, Nikolaev area; (*III*) Well 32, Astrakhan area; (*IV*) Well 13, Karachaganak area; (*V*) Well 5, Karachaganak area; (*VI*) Well 6, Noven'kaya area. 1, basin sediments; 2, sediments of wave zone shoal; 3, deposits of shallow shelf plane; 4, slope deposits; 5, lagoon sediments; 6, reef flat sediments. Localities of allochthonous assemblages: (a) synchronous; (b) asynchronous. Units: (vn) upper Visean Substage, Venev Horizon; (C_1s) Serpukhovian Stage; (C_2b_1) lower Bashkirian Substage; (krp) Krasnaya Polyana Horizon; (sk) Severnaya Kel'tma Horizon; (Prk) Kama Horizon; (C_2b_2) upper Bashkirian Substage; (C_2m) Moscovian Stage; (P_1a) Asselian Stage.

centrations of certain fossils, such as crinoids. This indicates that these allochthonous assemblages were formed in shoals, beaches, and bars under constant wave and current activity, in shallow water settings with a moderate hydrodynamic regime, as well as in highly energetic areas and zones of the sediment washout.

The alteration of groups of organisms producing carbonate sands reflected the evolutionary faunal changes. Crinoidal sands in the Caspian region were characteristic of the Famennian (Late Devonian), as well as of the Middle Carboniferous, foraminifer sands were characteristic of the Early and Middle Carboniferous, and fusulinid sands were typical of the Early Permian time.

The type of burial of organic remains in the facies of organogenic buildups depended on the hydrodynamic regime. Redeposited and damaged fossils were accumulated under highly energetic conditions, whereas autochthonous forms dominated in the environments with moderate water hydrodynamics. The allochthonous fossil assemblages of this facies are identified with difficulty. Thus, only one locality of asynchronous allochthonous assemblage from the upper Sakmarian reef facies was reliably distinguished in the studied area. In modern reefs, skeletons of reef constructors and reefophils primarily remain in the habitat area excluding those damaged by bioerosion, destroyed by waves, and transported to reef holes or down to the fore reef zone (James, 1983). Allochthonous skeletal remains, together with different sedimentary material, were recorded in reef holes. The coarse-grained skeletal sand, derived from the reef surface, occurs in large pores, whereas calcareous ooze with foraminifer shells and planktonic algae is usually found in smaller pores. In fossil reefs, allochthonous benthic assemblages were encountered in holes and fissures filled with syngenetic and epigenetic inner deposits (Burchette, 1981).

In the studied region, synchronous allochthonous assemblages were recorded in lower Bashkirian bioclastic limestones represented by the reef flat subfacies.

In the submarine slope facies, allochthonous benthic complexes occur in the lower Visean, Serpukhovian, Bashkirian, Sakmarian, and Artinskian sediments. Allochthonous assemblages are associated with bio- and lithoclastic breccia; brecciated, microbrecciated, and bioclastic limestones (grainstones); and with limestones with abundant matrix (flowstones). Bioclasts of local origin consist of skeletal fragments derived from a carbonate platform margin, or from a reef top and slopes.

In the studied region, synchronous allochthonous assemblages of the open shelf facies of the carbonate platform were recorded in Lower Devonian and Serpukhovian sediments. Microfacies containing allochthonous fossils are represented by microcrystalline limestones with rare intraclasts, and by organoclastic limestones (wackestones). It should be noted that in recent analogous facies, the mixed complexes dominate. The role of the autochthonous component increases under conditions of lower hydrodynamic energy and higher salinity. The allochthonous components consist of inhabitants of adjacent areas.

In the basin facies, allochthonous assemblages are known from Bashkirian microbedded limestones.

The revealed facies confinement of asynchronous and synchronous allochthonous assemblages suggests that their stratigraphic distribution can be substantially wider. Allochthonous assemblages can be found in similar facies at different stratigraphic levels.

FORMATION MODEL OF ALLOCHTHONOUS ASSEMBLAGES

The formation model of asynchronous allochthonous assemblages includes the complicated long-term interaction of sedimentary, biologic, and tectonic processes. Their formation could result from a series of events: breaks in sedimentation; sea level oscillations; discrete events of erosion and burial of fossils; weathering; dissolution of various sediments in accumulation areas and submarine paleorises with different amounts of benthic fossils resulting from biologic productivity and postmortal changes during the initial burial; and from the transportation and reworking of erosional products and their secondary burial.

The transportation of benthic fossils, both in rock fragments and after natural preparation, is controlled by a physical energy of the flux, namely, by wind, tidal, shore, and bottom currents, gravitational processes, as well as by the bottom inclination and resistance of different fossils to transportation and redeposition.

The formation of synchronous allochthonous assemblages is the result of combined sedimentation and biologic processes. It is characterized by: (1) a high productivity of benthic biota, (2) preparation, transportation, and reworking of benthic skeletons under an active or moderately active hydrologic regime prevailing during the accumulation, (3) lateral transportation of remains by short-term storms, (4) transportation of fossils by gravitational sliding, and (5) secondary burial. The influence of these factors differs in different paleoenvironments.

CONCLUSION

Accordingly, allochthonous benthic assemblages occur in the upper Paleozoic of the Caspian region considerably rarer than the autochthonous complexes. Two types of allochthonous assemblages, namely, asynchronous and synchronous with respect to enclosing sediments, are distinguished. The diagnostics of allochthonous assemblages is based on the analysis of paleontologic, paleoecologic, and sedimentologic criteria.

Allochthonous assemblages are of a wide stratigraphic range in the studied region. They are associated with Lower Devonian, lower Tournaisian, lower and upper Visean, Serpukhovian, lower Bashkirian, Asselian, Sakmarian, and Artinskian sediments.

Asynchronous allochthonous assemblages are most widespread at the base of transgressive cycles. The age difference between these assemblages and autochthonous complexes can be significant (when they belong to different systems or series), or insignificant (when they correspond to different stages). Asynchronous allochthonous assemblages were formed at the expense of the destruction of lithified or weakly consolidated sediments. Local paleorises were the main source of asynchronous allochthonous material.

The distribution of allochthonous assemblages is controlled not only by the age of the deposits, but by the facies as well. A considerably high water energy, which resulted in the destruction, reworking, and transportation of sediments, was the major factor responsible for the appearance of the studied assemblages. Synchronous allochthonous complexes are the most frequent in the facies of wave zone shoals, but also occur in the facies of lagoon, organogenic buildups, submarine slope, open shelf, and basins.

Reviewer A.S. Alekseev

REFERENCES

Bigon, M., Harner, J., and Traunsend, K., Ecology: Individuals, Populations, and Assemblages. Translated under the title *Ekologiya. Osobi, populyatsii i soobshchestva*, Moscow: Mir, 1989, vol. 1.

Burchette, P.T., European Devonian Reefs: a Review of Current Concepts and Models, *SEMP Special Publication*, 1981, no. 30, pp. 85–142.

Dunham, R.J., Classification of Carbonate Rocks According to Depositional Texture, in *Classification of Carbonate Rocks*, Ham, W.E., Ed., Mem. Am. Assoc. Petrol. Geol., 1962, vol. 1, pp. 108–121.

Dzhenchuraeva, A.V. and Soloshenko, I.I., On the Redeposition of Foraminifers in Upper Paleozoic Sediments from Eastern Altai, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1983, vol. 58, issue 2, pp. 75–82.

Flugel, E., *Microfacies Analysis of Limestones*, Berlin: Springer, 1982, vol. XIV.

Frankel, E., Recent Sedimentation in the Princess Charlotte Bay Area, Great Barrier Reef Province, *Proc. Int. Coral Reef Symp. 2nd*, Brisbane, 1974, pp. 355–369.

Goroshkova, V.A., Ovnatanova, N.S., and Zamilatskaya, T.K., On the Large Stratigraphic Unconformity in the Inner Flank Zone of the Caspian Syneclise, *Dokl. Akad. Nauk SSSR*, 1989, vol. 304, no. 3, pp. 676–679.

Hine, A.C., Lily Bank, Bahamas: History of an Active Oolite Sand Shoal, *J. Sediment. Petrol.*, 1977, vol. 47, no. 4, pp. 1554–1581.

Izotova, M.N., Lower Permian Fusulinid Zones from the Northern and Eastern Margins of the Caspian Depression, *Sov. Geol.*, 1985, no. 3, pp. 76–84.

James, N.P., Reef, in *Carbonate Depositional Environment*, Scholle, P.A., Bebout, D.G., and Moore, C.H., Eds., Tulsa: Am. Assoc. Petrol. Geol., 1983, pp. 346–440.

Klenina, L.N. and Ovnatanova, N.S., Facies Control of the Conodont Distribution in the Caspian Syneclise, *Izv. Akad. Nauk SSSR., Ser. Geol.*, 1986, no. 12, pp. 66–73.

Klenina, L.N. and Zaitseva, E.L., Conditions and Evidence of Benthic Fossil Redeposition in the Carboniferous of the

Caspian Depression, in *Predely tochnosti biostratigraficheskoi korrelyatsii* (Limits of Accuracy of Biostratigraphic Correlation), Leningrad: Nauka, 1995, pp. 132–140.

Kukhtinov, D.A., Alekseeva, V.I., and Kalmykova, A.G., Redeposition of Paleontologic Remains in the Permian and Triassic of the Caspian Depression and Its Possible Reasons, *Dokl. Akad. Nauk SSSR*, 1982, vol. 264, no. 3, pp. 672–674.

Kukhtinova, L.V. and Kukhtinov, D.A., Lower Permian Palynocomplexes in Terrigenous Sediments from the Eastern Caspian Depression, *Sov. Geol.*, 1990, no. 8, pp. 51–56.

Poyarkov, B.V., Degree of Influence of Foraminifer Redeposition on Precise Age Determinations of Siliceous–Carbonate Sediments from the South Fergana Depression, *Vopr. Mikropaleontol.*, 1973, no. 16, pp. 183–192.

Rauzer-Chernousova, D.M., *Foraminifery stratotipicheskogo razreza Sakmarskogo yarusa (r. Sakmara, Yuzhnyi Ural)* (Foraminifers from the Sakmarian Stratotype Section. The Sakmara River, Southern Urals), Moscow: Nauka, 1965.

Rozovskaya, S.E., Upper Carboniferous and Lower Permian Fusulinids of the Southern Urals, *Tr. Paleontol. Inst. Akad. Nauk SSSR*, 1952, issue 40, pp. 5–50.

Wilson, J.L., *Carbonate Facies in Geologic History*, 1977. Translated under the title *Karbonatnye fatsii v geologicheskoi istorii*, Moscow: Nedra, 1980.

Zamarenov, A.K., *Srednii i verkhnii Paleozoi vostochnogo i yugo-vostochnogo obramleniya Prikaspiiskoi vpadiny* (Middle and Upper Paleozoic of the Eastern and Southeastern Margin of the Caspian Depression), Leningrad: Nedra, 1970.

Zolotukhina, G.P. and Dan'shina, N.V., Structural Peculiarities of the Upper Paleozoic of Tengiz and Korolev Areas (Southeastern Caspian Depression), *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1992, no. 7, pp. 79–85.

Zolotukhina, G.P., Ketat, O.B., Dan'shina, N.V., and Aleshin, V.M., On Stratigraphy of the Upper Paleozoic Sediments from the Southeastern Caspian Depression (Tengiz Area), *Dokl. Akad. Nauk SSSR*, 1989, vol. 304, no. 4, pp. 933–937.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 341–361. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 31–52.
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Age of the Paramonovo Formation (the Albian of the Russian Platform) and Global Transgressive-Regressive Cycles of the Cretaceous

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Received March 19, 1994, in final form, January 20, 1994

Abstract—An assemblage of calcareous foraminifers (including the planktonic species), which was distinguished for the first time in the Paramonovo Formation of the Moscow region, suggests that the stratigraphic position of the unit corresponds to the terminal Albian (Vraconian), and likely, to basal Cenomanian. The sedimentological features and ecological type of fauna indicate that the initial basin depth of about 90–100 m decreased later to 40–50 m, as follows from the appearance of tempestites in the section. Sediments of the Paramonovo Formation mark the terminal Albian transgression spread over the Russian platform, when the sea level rose up to +240 m. Comparable amplitudes of sea-level rise were recorded only for the Oxfordian and late Santonian–early Campanian transgressions. The "Paramonovo event" of that high sea-level stand is also traceable in Western Europe.

Key words: *stratigraphy, upper Albian, foraminifers, radiolarians, spores, pollen, tempestites, transgression.*

Global eustatic transgressions and regressions are of interest to geoscientists, because they provide the sequence stratigraphic scale for the subdivision and correlation of Mesozoic–Cenozoic deposits, and are useful in hydrocarbon prospecting (Vail *et al.*, 1977; Haq *et al.*, 1987, 1988). However, the validity of this scale is doubtful, and recent ideas suggest that it be checked using sections of central cratonic areas, which are presumably the most stable in the tectonic aspect and preserved sedimentary cover, whose structure was defined by transgressive–regressive rhythms (Sahagian and Holland, 1991).

Among the regions promising in this respect is the East European (Russian) platform, one of the earth's largest Precambrian cratons. The Moscow syneclise in its center has a complex sedimentary cover of predominantly marine Jurassic (beginning with the Callovian) and Cretaceous deposits lying almost horizontally. This area is remote for a distance over 1000 km from the Paleozoic and younger fold belts of the Urals, Caucasus, and Carpathians. The Dnieper–Donets depression, which was actively subsiding during the Cretaceous–Paleogene time, is located 600 km southwest of Moscow.

In the central Russian platform, the most remarkable unit is the Paramonovo Formation of black and dark-gray silty clay, whose thickness range is 20–50 m. The Paramonovo Clay is widespread in the Klin–Dmitrov ridges north of Moscow, and extends from there to the northeast into the Yaroslavl and Vladimir regions (Olfer'ev, 1986). An outlayer of the Paramonovo Formation is known in the Teplyi Stan district of Moscow. This formation also covers wide areas in the Ryazan–Saratov depression and Ul'yanovsk region adjacent to the Volga River. Deposits similar in age and lithology were reported in addition from the Caspian depression (Sazonov and Sazonova, 1967).

The regional persistence of the Paramonovo Formation, its clayey composition, and abundant radiolarian fauna suggest that this unit was deposited at the time of a great transgression, when deep-water conditions were typical of the central Russian platform. Therefore, any precise dating of the Paramonovo Formation is of primary importance in calibrating the time scale of the transgressive–regressive rhythm. So far, reliable age data for the Paramonovo Clay are lacking, because in the stratotype area, these rocks are free of ammonites and foraminifers, the principal index fossils in the Cretaceous stratigraphic zonation.

For the entire period of investigation, the stratigraphic position of the Paramonovo Formation was thought to be corresponding to the upper Albian (Dobrov, 1932; Dan'shin, 1947; Gerasimov *et al.*, 1962; Gerasimov, 1971; Olfer'ev, 1986; Baraboshkin, 1991).

This conclusion was based upon two facts: (1) the Paramonovo Clay rests on the erosion surface at the top of distinctly middle Albian glauconite sands with phosphorites (the Gavrilkovo Formation in the scheme of Olfer'ev); (2) the eroded upper surface of the clay is overlain by sandy deposits with basal phosphorites

bearing Cenomanian ammonites. In addition, it was reported that in some localities of the Ryazan'–Saratov depression, the basal phosphorites of the formation enclose remains of the late Albian ammonites (Sazonov and Sazonova, 1967; Baraboshkin, 1991), and the radiolarian assemblage from this unit was earlier considered as correlative to the late Albian radiolarian fauna of the Crimea, although the latter has never been adequately described.

Thus, the Paramonovo Formation was dated with indirect data, which also offer the possibility to refer it to the Cenomanian, if its basal phosphorites are considered with the late Albian ammonites as redeposited from older beds now entirely eroded, which is typical of the Albian deposits in the Russian platform (Baraboshkin, 1991).

Below we present the results of our comprehensive micropaleontological analysis of the Paramonovo Formation in the Spas-Kamenka section, where we discovered for the first time calcareous benthonic and planktonic foraminifers, and also studied the diverse palynocomplexes. Our results afford the opportunity to correct the previous viewpoints on the age of this stratigraphic unit.

HISTORY OF INVESTIGATIONS

The first stratigraphic scheme for the Cretaceous deposits of the northern Moscow region (the Klin–Dmitrov ridges) was proposed by Sokolov (1913). He discriminated the unit of black clay and corresponding waterproof horizon, and named them after the Paramonova Village, where they were initially described (Sokolov, 1913; p. 9).

Later, when the toponym was modified, they became known under the name Paramonovo Clay (Milanovskii, 1926; Khimenkov, 1928; Rozanov, 1927; Dobrov, 1932), and this term is still used.

Prior to the mid-1980s, beginning with works of A.P. Pavlov and A.D. Arkhangel'skii, it was a common practice to distinguish in the Russian platform not the local formations, but rather, the West European stratigraphic units of the stage, or zone rank. The Paramonovo Clay was identified as the formation by Olfer'ev (1986), when the local lithostratigraphic scale of the central Russian platform was under elaboration.

As noted above, the Paramonovo Formation in the type locality (Dmitrov area) has not been characterized by micropaleontological data. Near the town of Yakhroma, Dobrov (1932) found only a doubtful fish-scale imprint and a lone remain of an arthropod.

Rozanov (1927, p. 143) believed that the Paramonovo Clay "can be conditionally correlated with the upper Albian (Vraconian Substage) clays of the Penza and Tambov provinces." The latter were referred to the Vraconian on the basis of the discovered remains of *Callihoplites vraconensis* Pictet et Campiche. Dobrov (1932; pp. 10–11) was of a similar opinion and wrote: "Because of their position between the middle Gault and Cenomanian, it is possible to consider the clays as a member of the former unit, or even of the upper Gault (Vraconian, Alb_5), especially in view of their distinctly gradual transition to overlying sands bearing the Cenomanian fauna at the top." The upper Albian level of the Paramonovo Clay was also accepted by Dan'shin (1947), Gerasimov *et al.* (1962), Sazonova and Sazonov (1967), and other scientists.

In the 1960s, it was discovered that in the Vladimir province (Vladimir and Kol'chugino localities), the Paramonovo Clay bears radiolarians. Aliev and Smirnova (1969), who briefly described these microfossils, pointed to their similarity with the Albian radiolarians from northeastern Azerbaijan. Aliev also distinguished some similar forms in the Albian deposits of the middle Volga region. The cited authors also listed several agglutinated foraminifers [*Ammobaculites agglutinans* (d'Orb.), *Gaudryinella* aff. *albica* Alekseeva, *Haplophragmoides nonioninoides* Reuss, and *Haplophragmium aequalis* Roemer] associated with radiolarians. Gerasimov (1971) repeated these data, and reported in addition that in its upper part, the units bear ammonites *Hoplites* sp. Baraboshkin and Mikhailova (1987), who revised this fauna, suggested that an ammonite form they studied, and the photograph of another specimen represent the *Pervinquieria* and *Hysteroceras* genera.

Baraboshkin and Mikhailova concluded that the lower part of the Paramonovo Formation is still of the middle Albian age, because the hiatus at its base is not very significant. Later, however, Baraboshkin (1991) repudiated this idea again in favor of the late Albian age of the entire Paramonovo Formation and sands of the Yakhroma Formation below the Cenomanian beds. Olfer'ev (1986, 1988) believed that in the type locality, the Paramonovo Formation tentatively is of the late Albian age, rests on the erosion surface, and bears foraminifers and radiolarians similar to the coeval forms from the Crimea. In his opinion, the Yakhroma Formation overlies the Paramonovo Clay with deep erosion and should be of the Cenomanian age.

In the volume of *"Stratigraphiya SSSR"* devoted to the Cretaceous, Gerasimov (1988) referred the Paramonovo Clay to the upper Albian, and pointed to the *Actinoceramus* cf. *sulcatoides* Saveliev which sometimes occurs in this unit.

The phosphorite horizon near the base of the overlying Lyaminskii Formation yielded Cenomanian ammonites and shark teeth (Milanovskii, 1926).

The review presented shows that previous paleontological data on the Paramonovo Formation are inadequate for the precise determination of the unit age. In addition, the cited authors controversially interpreted its relations with underlying and overlying deposits. In fact, the late Albian age of the formation is constrained by its stratigraphic position between the middle Albian and Cenomanian units.

SPAS-KAMENKA SECTION

The studied section is situated in the quarry, 25 km north of Moscow at the point of 56°09′N and 37°28′E (Fig. 1). The locality is removed for a distance of 11 km from the stratotype near the village of Paramonovo on the Volgusha River.

In the older southwestern part of the quarry, the lower beds are covered with landslides, and we studied the section of the northern wall, where clays were excavated in 1990.

The exposed section of the Spas-Kamenka quarry is 16.7 m thick, and displays three lithological members from the base upward (Fig. 2):

Member I: clay, black when wet, micaceous, with abundant marks of *Chondrites* life activity. The content of sandy fraction (>0.05 mm in grain size) is 5–7% of the rock volume. The apparent thickness is 1.7 m.

Member II: clay, dark gray or black when wet, micaceous. At the base, the unit is enriched in sandy-silty fraction including glauconite, and rests with the distinct ravinement on Member I. The green glauconite sand of the basal bed (0.1–0.15 m thick) penetrates down to the depth of few centimeters in the underlying clay. The content of sand fraction >0.05 mm is 17% at the base, and decreases to 2–5% at the top. In the lower part, clay beds are highly bioturbated, and some of them (lighter in color) are opoka-like. The thickness is 6.6 m.

Member III: clay, dark gray or black when wet, enclosing abundant lenticules and laminae (0.05–0.1 m thick) of microlaminated clayey to sandy silts. At the base, there is a persistent bed (up to 10 cm) of greenish-gray silt with yellow jarosite blooms. This bed is saturated with water and clearly seen as a dark stripe in the exposure. At the level of 1.1 m above the base, there is a discontinuous interbed (1–6 cm) of cross-laminated clayey silt, where laminae are 0.5–2 mm thick. The cross-bedding is accentuated by very fine black gouges. The cross-bedded series enclose fine rounded pebbles of the black clay. In the lower 4.5 m of the member, the clay contains 3–7% of clastic fraction (>0.05 mm in grain size), and 10–20% above this level. Lenticules and laminae of microlaminated sandy silt can be considered as tempestites, implying a lesser depth of sedimentation commensurable with the wave base. The thickness is 8.4 m.

In the Spas-Kamenka quarry, the base of the Paramonovo Formation is unexposed. The total thickness of this three-member unit normally is about 30 m in the region under discussion. Its basal beds consist of clayey glauconite sand locally enclosing interbeds of clay and phosphorite nodules. These beds are 2–5 m thick. The middle interval of the unit is composed of dark-gray and black silty clay (20–25 m). The formation is crowned by dark-gray, often mottled clayey sand (up to 4–5 m thick). Consequently, the studied section of the Paramonovo Formation comprises its middle interval almost in full measure, because the drilling data

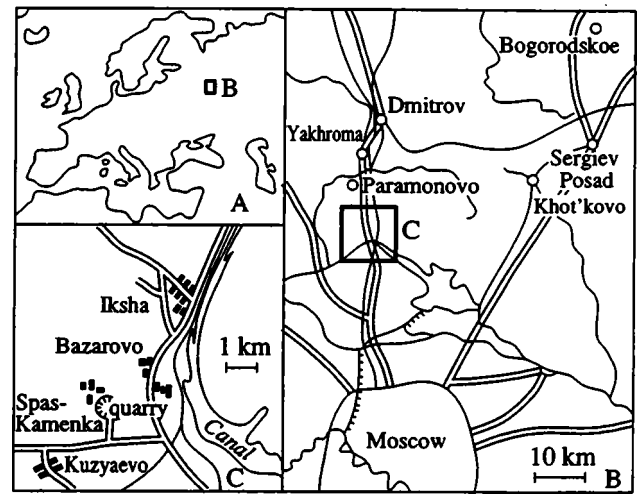

Fig 1. Geographic position of the Spas-Kamenka section.

(Bizyaeva and Prints, 1974) indicate that the clay beds of the Spas-Kamenka quarry are 18 m in total thickness.

MATERIALS AND METHODS

We studied 36 samples collected from the section with intervals of 0.5–1 m between them. Sample fragments, each 100 g in weight, were disintegrated by boiling them in the sodium bicarbonate solution to extract microfossils from their fractions >0.05 mm in grain size. The mass of Sample SK-27, the most rich in calcareous foraminifer shells, was about 0.5 kg. Palynoflora was separated from 20 samples by the standard method and CdJ + KJ heavy liquid. With the exception of very poor extracted assemblages, not less than 200 spore–pollen grains were counted in other cases. The foraminifer collection is stored under no. 266 at the Department of Paleontology, Geological Faculty of Moscow State University. The collection of radiolarians is reserved at the Geological Institute, Russian Academy of Sciences.

FORAMINIFERS

Throughout the studied section, clays of the Paramonovo Formation yielded foraminifers, though not very abundant (usually a dozen of specimens per a sample). The samples mostly contain a stable species set of agglutinated foraminifers of *Reophax, Ammobaculites, Haplophragmoides, Lagenammina*, and *Trochammina* genera. Their tests usually are of a very good preservation state, although sometimes the tests are deformed because of the clay consolidation. Within the interval of 1.4–4.0 m above the Member II base, Samples SK-25–SK-28 contain in addition to agglutinated, calcareous benthonic (mostly nodosariids) and planktonic forms. Calcareous foraminifers are poorly preserved; their

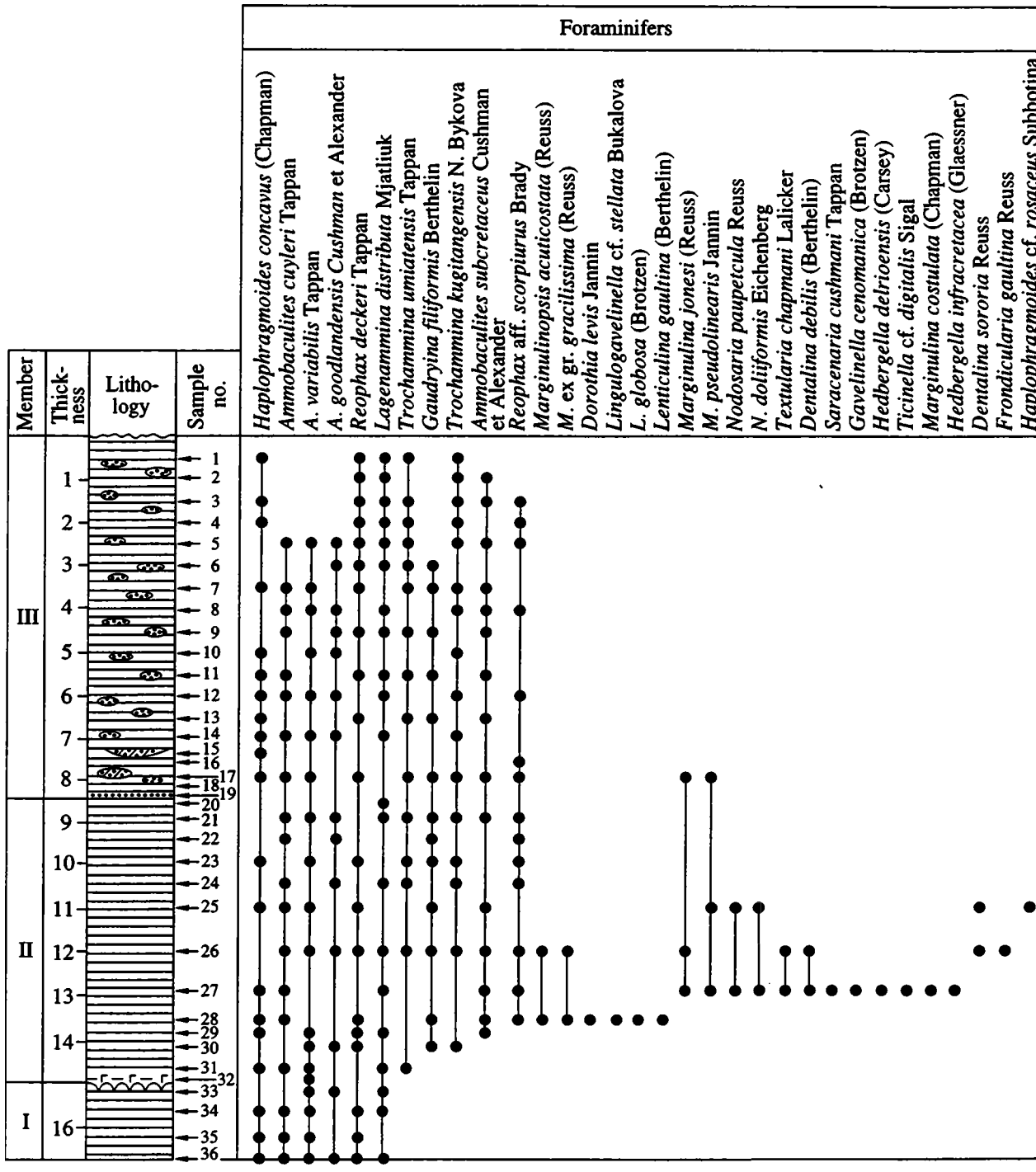

Fig. 2. Distribution of foraminifers, spores, and pollen through the Spas-Kamenka section of the Paramonovo Formation: 1, clay; 2, silt; 3, glauconite; 4, cross-bedding.

tests are corroded and somewhat recrystallized. These forms are considerably less abundant than agglutinated foraminifers. Very rare specimens of benthonic calcareous foraminifers were also distinguished in the basal part of Member III (Sample SK-17). In addition, Sample SK-11 (2.8 m above Member III base) yielded an indeterminable specimen of a calcareous foraminifer.

We distinguished a total of 32 foraminifer species of 19 genera, including two planktonic taxa of the latter rank. The studied benthonic foraminifers include eight genera with agglutinated shells mostly consisting of quartz grains; calcareous shells are typical of nine genera, and basic species of both types are pictured in Plates I–III.

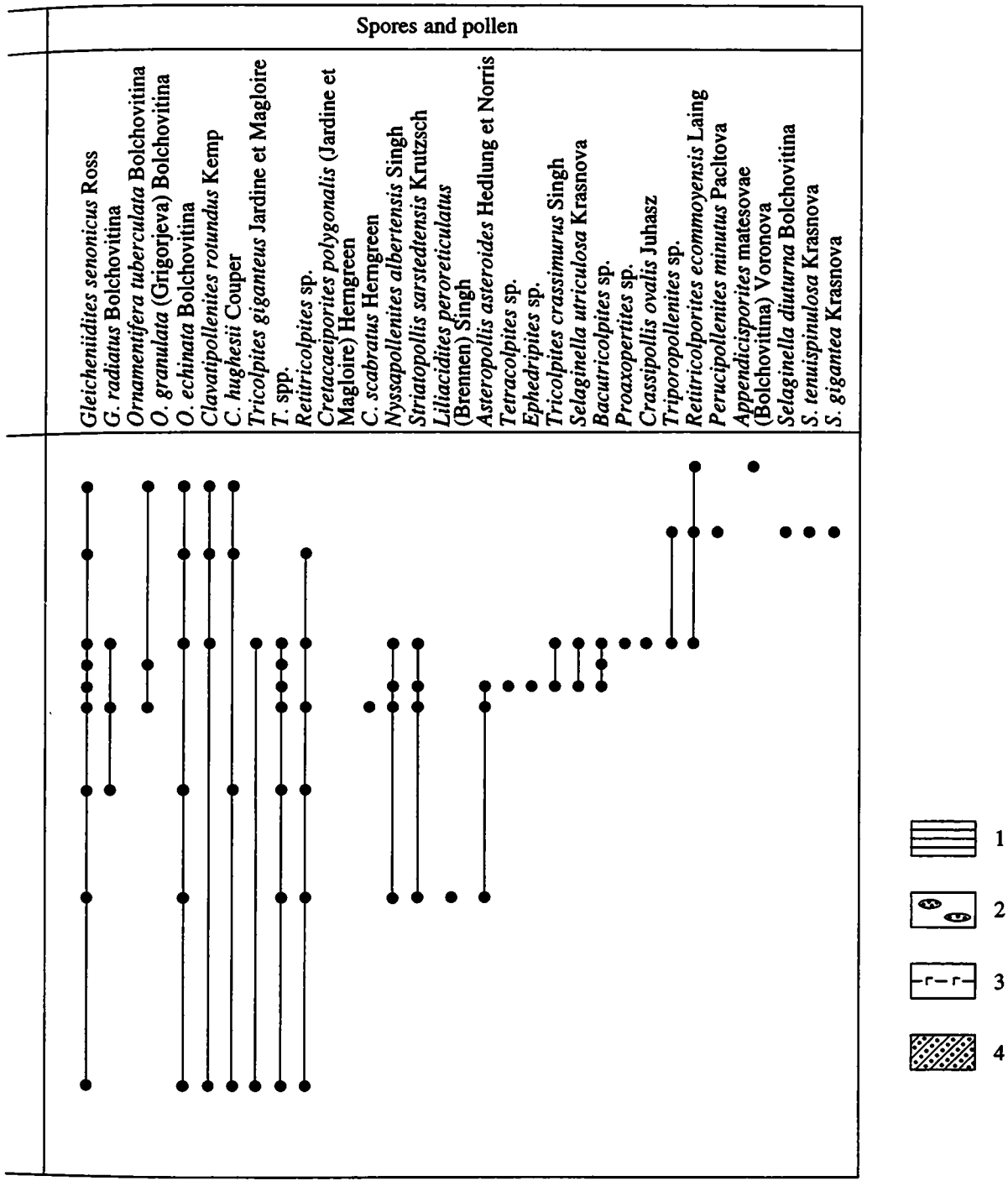

Fig. 2. (Contd.)

Analyzing the vertical distribution of foraminifers in the Spas-Kamenka section, we identified three successive assemblages (Fig. 2).

Assemblage I is confined to the lower member, and includes a few agglutinated forms, which also occur in overlying members. These are *Ammobaculites variabilis Tappan, A. cuyleri Tappan, A. goodlandensis* Cushman, *Reophax deckeri Tappan, Lagenammina distributa Mjatliuk,* and *Haplophragmoides concavus* (Chapman). The first two species and *Reophax deckeri* Tappan were originally described from the upper Albian Grayson Formation of northern Texas (Tappan, 1940). However, *R. deckeri* was also reported from different horizons of the Albian and lower Cenomanian deposits of Texas and Oklahoma (Frizzel, 1954; Tappan, 1943). *Ammobaculites goodlandensis* Cushman et

Alexander occurs in the Albian deposits of North America (Cushman and Alexander, 1930; Loeblich and Tappan, 1949), and such species as *Haplophragmoides concavus* (Chapman) is widespread in Aptian and Albian sediments.

Assemblage II was distinguished in Member II and at the base of Member III. In addition to agglutinated foraminifers typical of the previous assemblage, the last one includes *Trochammina umiatensis* Tappan, *T. kugitangensis* N. Bykova, *Gaudryina filiformis* (Berthelin), *Ammobaculites subcretaceous* Cushman et Alexander, and *Reophax* aff. *scorpiurus* Brady. The first species among these forms is typical of the middle–upper Albian Grandstan Formation of Alaska (Tappan, 1957). It is also widespread in the middle Aptian–middle Albian deposits of the East European platform, as follows from the data of Mjatliuk and Vassilenko (1988), who also reported that *Trochammina kugitangensis* N. Bykova is characteristic of the upper Albian–lower Cenomanian deposits occurring in Mangyshlak, Ustyurt, and eastern Turkmenistan. The species that was classified as *Reophax* aff. *scorpiurus* Brady is identical to *R. scorpiurus* Brady from the upper Albian type section in the Parisian basin (Magniez-Jannin, 1975). However, this form was first reported from the recent deposits, and its name should be used in terms of the open nomenclature to designate the Cretaceous form. The assemblage in question is most diverse within the interval of 1.4–4.0 m above the base of Member II, where the listed agglutinated foraminifers occur together with calcareous forms. Sample SK-27 from this interval yielded 17 species. It is also remarkable that two samples (SK-26 and SK-27) showed the presence of *Textularia chapmani* Lalicker, the species known from Albian strata of the Netherlands (Ten Dam, 1950) and Poland (Alexandrowiz, 1978), and also typical of the upper Albian (Vraconian) beds in the Parisian basin (Magniez-Jannin, 1975). Among the calcareous foraminifers, the most abundant are *Dentalina inconstans* Jannin, *D. debilis* (Berthelin), *D. sororia* Reuss, *Marginulina jonesi* (Reuss), *M. pseudolinearis* Jannin, and *Marginulinopsis acuticostata* (Reuss). These species were reported from the different Albian levels of Western Europe, and many of them occur in the Vraconian deposits of the Parisian basin. Another group of rare species includes *Trochammina* cf. *bulloides* Tairov and *Lingulogravilinella* cf. *stellata* Bukalova. These forms were distinguished in the Albian beds of Azerbaijan and northern Ciscaucasia, respectively (Bukalova, 1958). Two other representatives of this group, *Dorothia levis* Jannin and *Marginulinopsis acuticostata restricta* (Jannin), occur in the upper Albian strata of the Parisian basin (Magniez-Jannin, 1975).

In addition to the aforementioned species, the most diverse assemblage from Sample SK-27 includes *Saracenaria cushmani* Tappan occurring in the upper Albian beds of Texas, *Gaudryina* sp., and *Gavelinella cenomanica* (Brotzen). The last species is most remarkable, because it is widespread in the lower Cenomanian beds of the East European platform, Crimea, northern Caucasus, and Texas, and has been also identified in the Albian–Cenomanian deposits of Sweden (Brotzen, 1945; Vassilenko, 1961). Data of the Deep Sea Drilling Project were used to distinguish the *Gavelinella cenomanica* Zone within the Albian–Cenomanian oceanic sections (Riegraf and Luterbacher, 1989). This zone is also marked by presence of *Lingulogavelinella globosa* (Brotzen), the species widespread in the upper Albian–lower Cenomanian beds of the Crimea, northern Caucasus, and Mangyshlak (Vassilenko, 1961).

We encountered very rare and poorly preserved specimens of planktonic foraminifers only in Sample SK-27. Among them, we identified *Hedbergella infracretacea* (Glaessner), the species of global distribution and most typical of the upper Albian–lower Cenomanian level. Two other distinguished forms are *Hedbergella delrioensis* (Carsey), which is known from the uppermost Albian and lower Cenomanian deposits, and *Ticinella* cf. *digitalis* Sigal. *Ticinella digitalis* Sigal was originally described from the Vraconian strata of Madagascar (Sigal, 1966), and then distinguished in the upper Albian beds of Bavaria and the North Atlantic Ocean (Risch, 1971; Gorbachik, 1980).

The most significant form among planktonic foraminifers is *Hedbergella delrioensis* (Carsey). Its stratigraphic range has been controversially interpreted in available publications. Many researchers believe that the *Rotalipora ticinensis* Zone of the upper Albian corresponds to the first appearance level of this species, and place the upper limit of the species distribution at the Albian–Cenomanian boundary (Moullade, 1966; Risch, 1971; Magniez-Jannin, 1975). Others (Krasheninnikov and Basov, 1985; Caron, 1985) report on its distribution through the entire Albian and, perhaps, the upper Aptian intervals. The last opinion is based on the unreasonably argued idea to consider the stratigraphic range of *Hedbergella delrioensis* (Carsey) as that also comprising the distribution levels of the Aptian–Albian form *Hedbergella aptica* (Agalarova) and some other species. We favor the first viewpoint.

In addition, it is remarkable that *Hedbergella delrioensis* (Carsey) seems to be a synonym of *Hedbergella caspia* (Vassilenko). Both species are very similar in morphology and show identical intervals of stratigraphic distribution; however, the first name is popular in Western European publications, while the second one frequently occurs in Russian works.

Assemblage III is distributed through the main part of Member III. It consists of agglutinated foraminifers only, among which we recognized all of the species of Assemblage II with the exception of *Dorothia levis* Jannin, *Textularia chapmani* Lalicker, and *Haplophragmoides* cf. *rosaceus* Subbotina. The assemblage appears to be considerably impoverished in the uppermost part of Member III (the interval of 6.4–8.4 m above its base; Samples SK-1 to SK-4). The last occur-

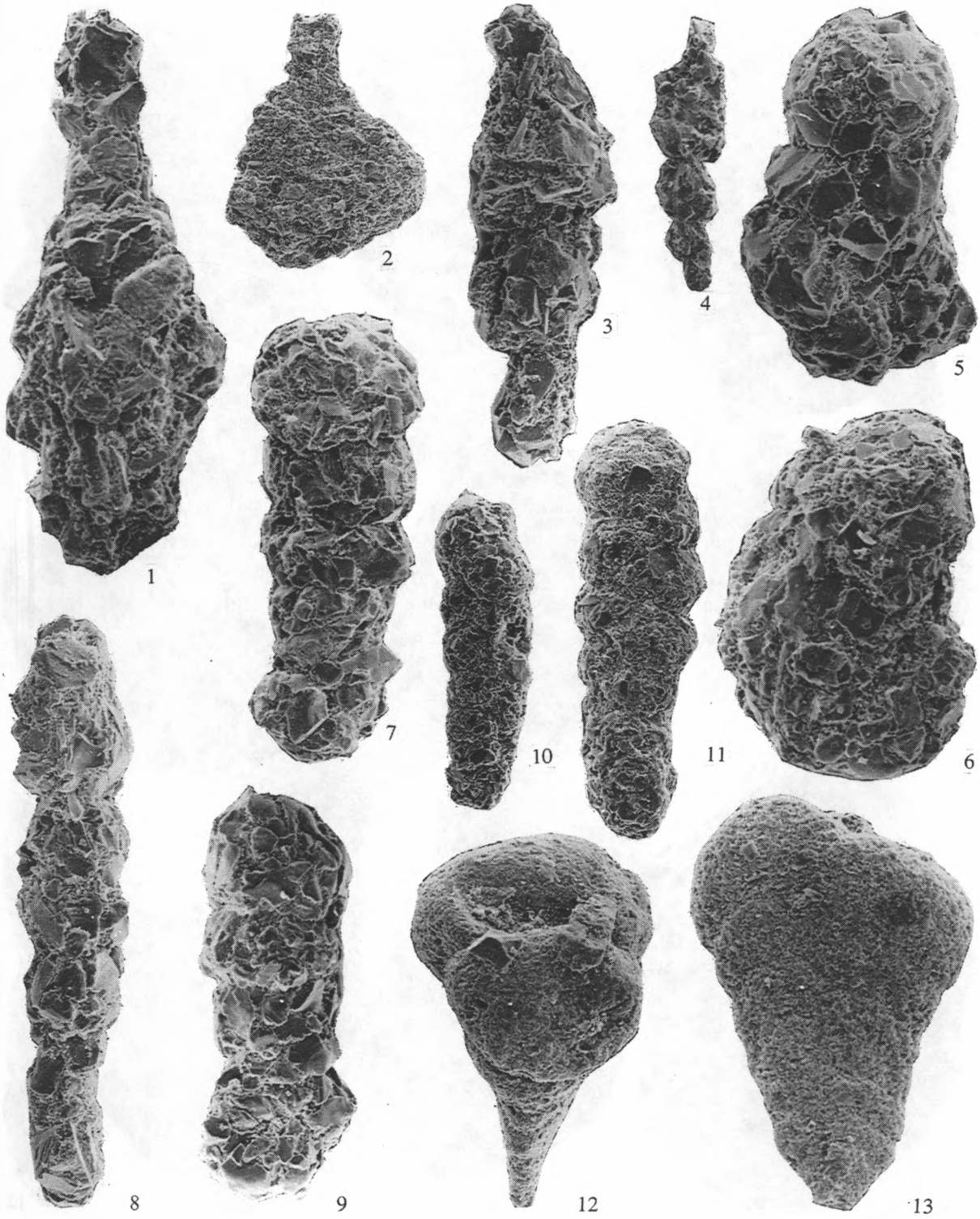

Plate I. Foraminifers from the Paramonovo Formation, the Spas-Kamenka section (Moscow region).

Key: (1, 2) *Lagenammina distributa* Mjatliuk: (1) specimen no. 266/1, peripheral side, ×100; (2) specimen no. 266/2, side view, ×47; (3, 4) *Reophax deckeri* Tappan: (3) specimen no. 266/3, ×100; (4) specimen no. 266/4, side view, ×50; (5, 6) *Ammobaculites goodlandensis* Cushman et Alexander: (5) specimen no. 266/69, ×150; (6) specimen no. 266/18, side view, ×100; (7–9) *Ammobaculites subcretaceous* Cushman et Alexander: (7) specimen no. 266/62, side view, ×100; (8) specimen no. 266/5, peripheral side, ×100; (9) specimen no. 266/66, side view, ×100; (10, 11) *Gaudryina filiformis* Berthelin: (10) specimen no. 266/60, ×100; (11) specimen no. 266/61, ×100; (12, 13) *Textularia chapmani* Lalicker: (12) specimen no. 266/14, side view, ×100; (13) specimen no. 266/16, aperture view, ×100.

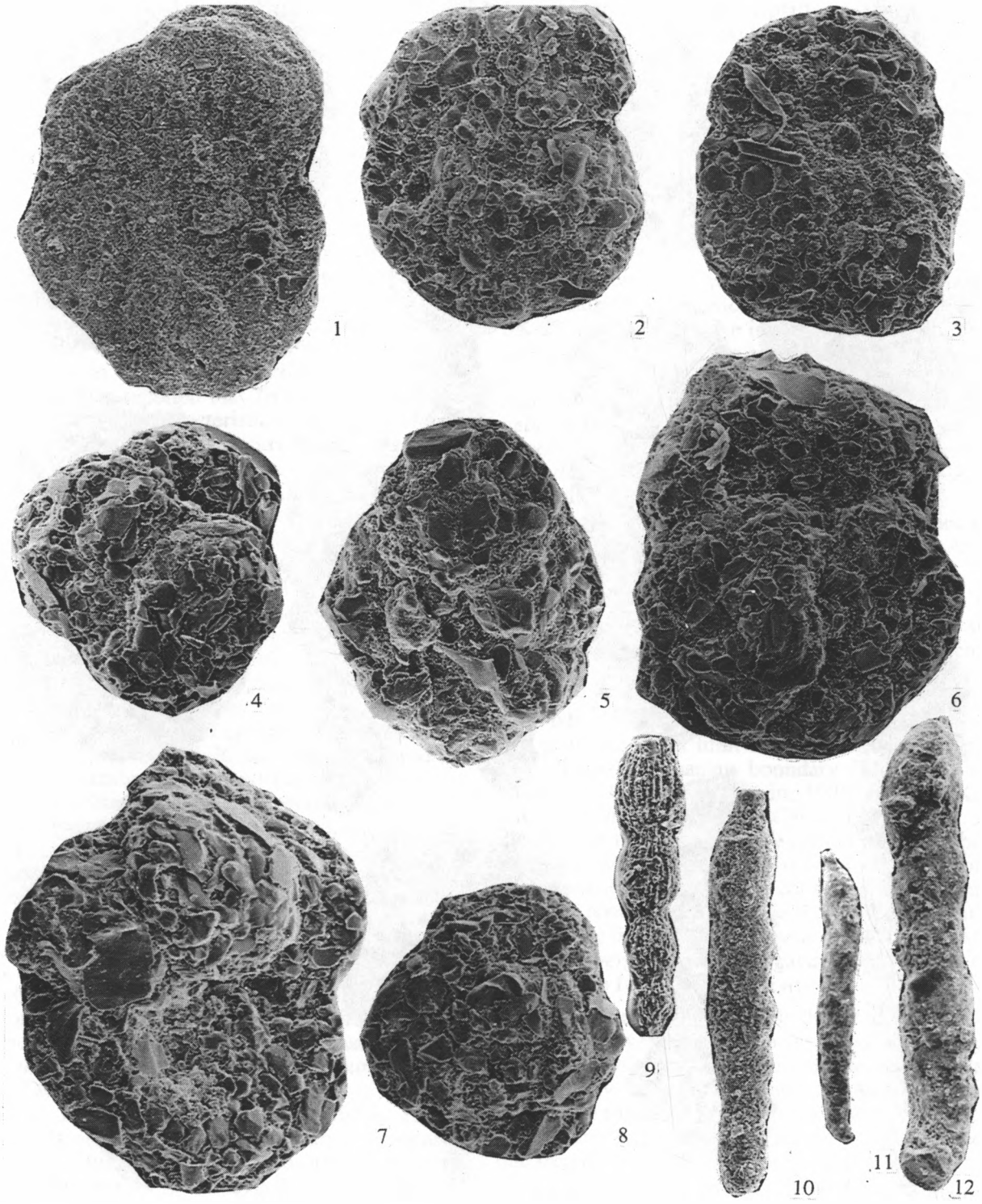

Plate II. Foraminifers from the Paramonovo Formation, the Spas-Kamenka section (Moscow region).

(1) *Haplophragmoides* cf. *rosaceus* Subbotina, specimen no. 266/8, side view, ×100; (2, 3, 5) *Trochammina kugitangensis* N. Bykova: (2) specimen no. 266/56, spiral side, ×100; (3) specimen no. 266/7, umbilical side, ×100; (5) specimen no. 266/72, spiral side, ×100; (4, 8) *Trochammina umiatensis* Tappan: (4) specimen no. 266/71, ×100; (8) specimen no. 266/57, spiral side, ×100; (6, 7) *Trochammina* aff. *umiatensis* Tappan: (6) specimen no. 266/55, spiral side, ×100; (7) specimen no. 266/70, umbilical side, ×100; (9) *Nodosaria paupercula* Reuss, specimen no. 266/10, ×100; (10) *Nodosaria doliiformis* Eichenberg, specimen no. 266/ 12, ×120; (11, 12) *Marginulina pseudolinearis* Jannin: (11) specimen no. 266/34, ×53; (12) specimen no. 266/37, ×100.

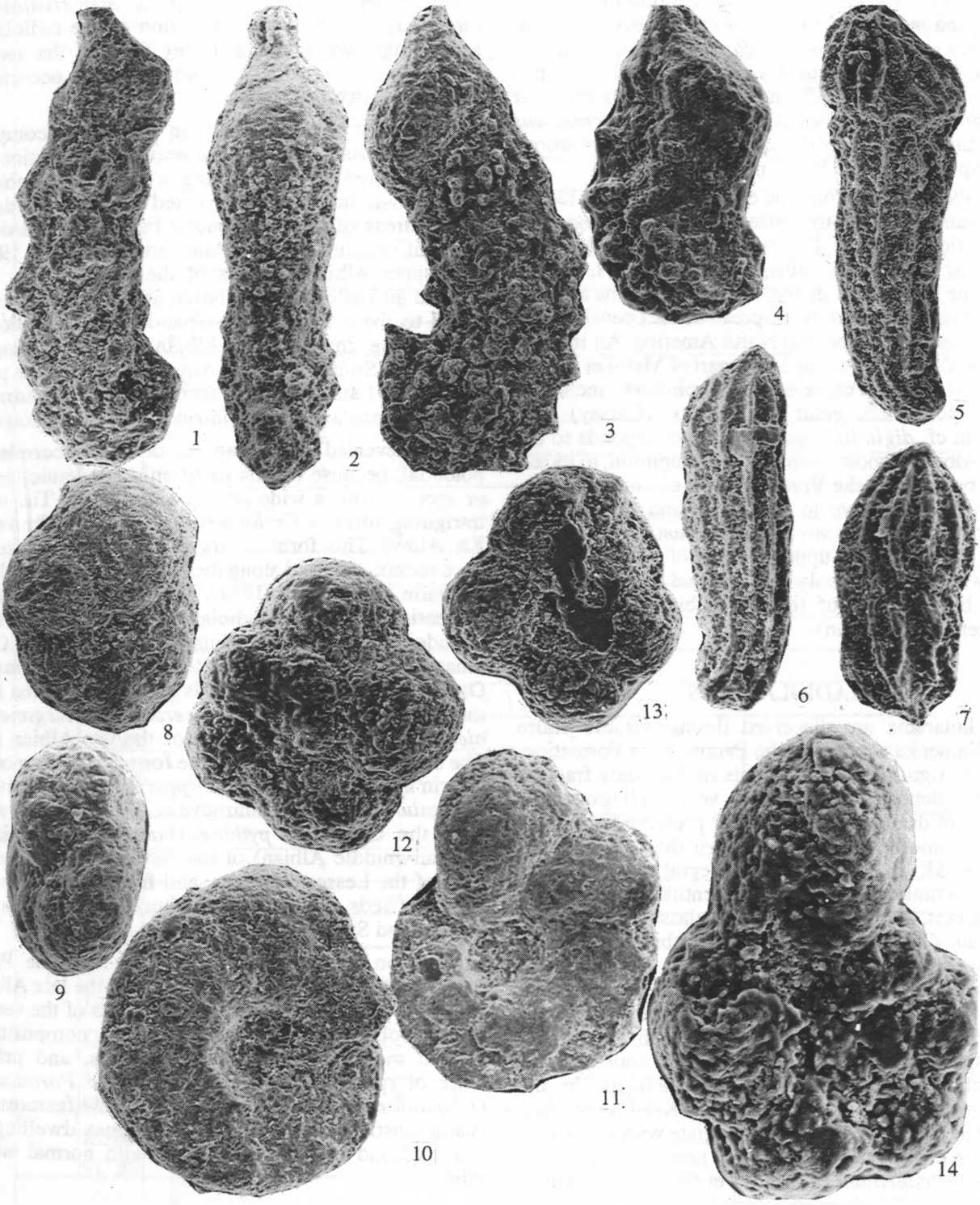

Plate III. Foraminifers from the Paramonovo Formation, the Spas-Kamenka section (Moscow region).

(1–4) *Marginulinopsis* ex gr. *gracillissima* (Reuss): (1) specimen no. 266/25, side view, ×100; (2) specimen no. 266/33, peripheral side, ×100; (3) specimen no. 266/24, side view, ×100; (4) specimen no. 266/39, side view of a young shell, ×100; (5, 6) *Marginulinopsis acuticostata* Reuss: (5) specimen no. 266/27, side view, ×100; (6) specimen no. 266/40, side view, ×100; (7) *Marginulina jonesi* (Reuss), specimen no. 266/296, side view, ×100; (8, 9) *Lingulogavelinella* cf. *stellata* Bukalova: (8) specimen no. 266/44, spiral side, ×120; (9) specimen no. 266/46, peripheral side, ×120; (10, 11) *Gavelinella cenomanica* (Brotzen): (10) specimen no. 266/20, ×100; (11) specimen no. 266/17, umbilical side, ×100; (12, 13) *Hedbergella delrioensis* (Carsey): (12) specimen no. 266/49, spiral side, ×150; (13) specimen no. 266/53, broken shell from umbilical side, ×150; (14) *Ticinella* cf. *digitalis* Sigal, specimen no. 266/54, spiral side, ×330.

rence level of *Gaudryina filiformis* (Berthelin) was distinguished in Sample SK-6. For three species (*Ammobaculites cuyleri* Tappan, *A. variabilis* Tappan, and *A. goodlandensis* Cushman et Alexander), this level corresponds to Sample SK-5, and such species as *Reophax* aff. *scorpiurus* Brady and *Ammobaculites subcretaceus* Cushman et Alexander do not occur above the positions of samples SK-3 and SK-2, respectively.

Analyzing the taxonomic composition of the identified foraminifers and their stratigraphic and geographic distribution, we arrived at several conclusions. Member I, bearing the first assemblage of foraminifers, is most likely the late Albian in age, because the same agglutinated foraminifers are widespread in the coeval deposits of Western Europe and North America. An interval of Assemblage II, i.e., the lower part of Member II with the most diverse calcareous foraminifers, including planktonic *Hedbergella delrioensis* (Carsey) and *Ticinella* cf. *digitalis* Sigal, clearly corresponds to the upper Albian, almost definitely in our opinion, to its terminal period, i.e., the Vraconian. We consider the Vraconian as correlative to the *Stoliczkaia dispar* Zone comprising the *Mortoniceras rostratum* and *M. perinflatum* subzones. The upper member of the studied section, which yielded only agglutinated foraminifers of Assemblage III, is of the late Albian age, or even younger (Cenomanian).

RADIOLARIANS

Radiolarians are dispersed throughout the entire Spas-Kamenka section of the Paramonovo Formation. They are significant components of the sandy fraction in these deposits, though not very well preserved because of dissolution. The better preserved and most diverse radiolarians were distinguished in Samples SK-22 to SK-28, i.e., within the interval bearing calcareous foraminifers. In total, we identified 12 forms of radiolarians, many of which can be classified at the species rank (Table 1). The four most abundant species (*Cromyodruppa concentrica* Lipman, *Orbiculiforma persenex* Pessagno, *Porodiscus kavilkinensis* Kh. Aliev, and *P. inflatus* Smirnova et Kh. Aliev) are spread throughout the section. The distribution patterns of other specimens allowed us to recognize four intervals with characteristic radiolarian assemblages. In the lower part of the section (Samples SK-29 to SK-36), the listed background species associate with *Holocryptacanium* cf. *barbui* Dumitrica and rare *Spongopyle* sp., *Crolanium cuneatum* (Smirnova et Kh. Aliev), *Pentinastrum* sp., and *Dictyomitra ferosia angusta* Smirnova. Upsection (Samples SK-20 to SK-28), the assemblage includes three additional forms: *Histiastrum latum* Lipman, and more rare *Crucella* sp. 1 and *Crucella* sp. 2. Sample SK-19 is free of radiolarians.

Such forms as *Crolanium cuneatum* (Smirnova et Kh. Aliev), *Pentinastrum* sp., and *Crucella* species do not occur in the lower beds of Member III (Samples SK-8 to SK-18), which still retain specimens of *Holocryptacanium* cf. *barbui* Dumitrica and *Histiastrum latum* Lipman. A further depletion of the radiolarian assemblage was noted in upper beds of the section (Samples SK-1 to SK-7), above the last occurrence level of the two last species.

Despite some fluctuations in taxonomic composition, radiolarians from the Paramonovo Formation can be considered as representing a single assemblage. Similar radiolarians were reported from eastern distribution areas of the Paramonovo Formation (Vladimir and Kol'chugino districts; Aliev and Smirnova, 1969). The upper Albian deposits of the Volga region also yielded an analogous radiolarian assemblage. As compared to these regions, the Spas-Kamenka section is poor in the characteristic Albian species *Crolanium cuneatum* (Smirnova et Kh. Aliev), but shows the presence of the Late Cretaceous forms, such as *Histiastrum latum* Lipman and *Orbiculiforma persenex* Pessagno.

The revealed assemblage is of a low correlation potential, because it consists of either endemic forms, or species with a wide stratigraphic range. The most intriguing form is *Crolanium cuneatum* (Smirnova et Kh. Aliev). This form occurs in the Cretaceous reference section exposed along the Naiba River of southern Sakhalin (Kazintsova, 1987). Member II of the Naiba Formation yielded a radiolarian assemblage, which includes, according to Kazintsova's data, several Cenomanian species in addition to *Crolanium cuneatum*. On the basis of inocerams, this unit of the Naiba Formation is referred to the *Inoceramus dunveganensis aiensis* Beds considered to be of the late Albian age. The overlying Member III of the formation is Cenomanian in age (the *Inoceramus crippsi* Zone). In addition, *Crolanium cuneatum* (Smirnova et Aliev) was reported from the *Crolanium pythiae–Thanarla conica* Beds (Aptian–middle Albian) of the Sevan–Akera tectonic zone of the Lesser Caucasus, and from the *Crolanium pythiae* Beds (Barremian–Aptian) of the Koryak Upland and Sakhalin (Vishnevskaya, 1991).

Data on radiolarians also suggest that the Paramonovo Formation is most probably of the late Albian age. The radiolarian assemblage proper is of the neritic type, as follows from the poor taxonomic composition, discoid morphotypes among dominants, and prevalence of species representing the genera *Porodiscus, Orbiculiforma*, and *Cromyodruppa*. These features are characteristic of radiolarian assemblages dwelling in marginal and epicontinental seas with normal water salinity.

OSTRACODES

In addition to planktonic and benthonic foraminifers, we extracted about 20 poorly preserved shells and dismembered valves of ostracodes from Sample SK-27. Yu.N. Andreev, who studied this material, identified among them *Rehacythereis* sp., *Protocytherinae* gen. et sp. indet., *Cytherella* sp., *Schulderidea* sp. indet.,

Table 1. Distribution of radiolarians though the section of the Paramonovo Formation

| Species | Sample numbers | | |

Amphicytherina sp., *Centrocythere* (?) sp., and *Quasihermanites* sp. In his opinion, the genera of this assemblage are typical of the Lower Cretaceous, or more precisely, of the Aptian–Albian deposits of Western Europe.

SPORES AND POLLEN

Palynological data were obtained for members II and III of the Spas-Kamenka section. Its Member I is extremely poor in palynofossils.

Palynospectra of the Paramonovo Formation contain diverse pollen of angiosperms and display a low abundance of conifers. The content of angiosperm pollen varies from 4 to 28%, and is at a minimum (4–5%) in the lower part of Member III; at the base of Member II, this pollen constitutes 24% of the counted grains.

The revealed palynospectra represent two main palynocomplexes primarily differing in their composition of angiosperm pollen (Fig. 2).

Palynocomplex I comprises spectra distinguished in Member II and at the base of Member III (Samples SK-18, SK-24, and SK-31). Palynoflora is not very abundant in these samples, and the calculations presented below to illustrate the proportions of different taxa are of questionable accuracy, since they are based in this case on only 50 grains counted per sample.

Fern spores (mostly of Gleicheniaceae) represent 38–62% of this palynocomlex, thus being dominant among other pollen and spore grains. The distinguished species include *Ornamentifera echinata* Bolch., *O. granulata* (Grig.) Bolch., O. tuberculata Bolch., *Gleicheniidites radiatus* Bolch., and *G. bolchovitinae* Dr. The most abundant grains belong to *Gleicheniidites senonicus* Ross, *G. laetus* Bolch., *G. umbonatus* Bolch., and *G. apilobatus* Bren.; Less abundant are *Plicifera delicata* Bolch. and *Clavifera triplex* Bolch. Many spores of Gleicheniaceae are stunted and small. Spores of Schizaeaceae (1–4%) are represented by fragmentary envelopes, and we met difficulties in attemping their precise identification. Nevertheless, it is possible to state that specimens of genera *Cicatricosisporites, Appendicisporites*, and *Schizaeosporites* are present among them. Other taxa, such as *Sphagnumsporites* sp., *Foveosporites canalis* Balme, *Cyathidites minor* Coup., and others are exemplified by single grains.

Scant grains of gymnosperm pollen were classified as *Monosulcites* sp., *Glyptostrobus* sp., *Sequoia* sp., *Taxodiaceaepollenites hiatus* (Pot.) Kemp., Cupressaceae, *Sciadopitispollenites* sp., and the bisaccate conifers of *Pinuspollenites* and *Podocarpidites* genera.

The content of angiosperm pollen varies in spectra from 4 to 24%. The monocolpate grains represent *Clavatipollenites hughesii* (Coup.) Kemp., *C. rotundus* Kemp., and *Liliacidites peroreticulatus* (Bran.) Singh, which were reported from the Barremian–lower Aptian deposits of southern England (Couper, 1958), Hungary (Juhasz, 1983), Ukraine (Voronova, 1991), and other regions.

Tricolpate pollen, usually typical of the late Aptian, is rather diverse, and includes specimens of *Tricolpites sagax* Nor., *T. giganteus* Jard. et Magl., *T.* sp. A., *Retitricolpites vulgaris* Pierce, and *R.* sp. A. The identified peculiar pollen of *Asteropollis asteroides* Hedl. et Nor. is known to occur in the lower–middle Albian and higher. Finally, it is important that the complex in question also contains such pollen taxa as *Cretaceaiporites polygonalis* (Jard. et Magl.) Hern., *Striatopollis sarstedtensis* Krutzsch, and *Nyssapollites* sp., which were widespread in England and North America beginning in the late Albian time.

Palynocomplex II is of a different type, and comprises six palynospectra from the upper part of Member III (Samples SK-2, SK-5, SK-9–SK-12). Rocks here are richer in pollen, whose diversity is considerably higher than in the older Palynocomplex I.

Gleicheniaceae still represent the dominant fraction here (35–63%), and in addition to the taxa distinguished in the lower palynocomlex, they include *Ornamentifera baculata* Singh, *O. conspicienda* Bolch, *Gleicheniidites minor* Dr., *G. carinatus* Bolch., *G. delcourti* Dr., and *G. saparicus* Juhasz. Spores of Schizaeaceae are more abundant (2–7%) and diverse than below. In addition to the spores of a wide stratigraphic range, this group also includes specimens mostly typical of the Albian–Cenomanian strata. Abundant spores of the first type belong to *Cicatricosisporites minutaestriatus* (Bolch.) Poc., *C. pseudoauriferus* Bolch., and *C. australiensis* Coup.; and specimens of *C. perforatus* (Bar., Nem. et Kondr.) Singh, *Appendicisporites matesovae* (Bolch.) Vor., *A. insignis* (Mark.) Chl., and *Corniculatisporites* sp. represent the second type. The less abundant taxa of a wide stratigraphic range are *Sphagnumsporites* spp., *Foveosporites canalis* Balme, *Densoisporites velatus* Weyl. et Krieg., *Laevigatosporites ovatus* Weyl. et Webst., *Cyathidites minor* Coup., *C. australis* Coup., *Matonisporites* sp., and *Leptolepidites verrucatus* Coup. Rare taxa typical of the Albian–Cenomanian deposits include *Coronatispora valdensis* (Coup.), Dett., *Rouseisporites reticulatus* Poc., *Murospora florida* Vor., and *Polypodiisporites* cf. *cenomanicus* Singh.

It is remarkable that Palynocomplex II also includes spores of Lycopodiophyta, such as *Selaginella tenuispinulosa* Krasn., *S. diuturna* Bolch., and *S. gigantea* Krasn. Originally, Krasnova (1961) discovered these species in the Cenomanian–Senonian deposits of West Siberia, and later, two of them (*S. tenuispinulosa* and *S. diuturna*) were distinguished in the Cenomanian–Coniacian sediments of Azerbaijan (Smirnova and Aliev, 1984).

Many spectra of this palynocomplex display a minor amount (1–3%) of gymnosperm pollen, whose content grows (up to 20%) only in Sample SK-11 at the expense of bisaccate pollen, generally representing

genera *Cedripites* and *Podocarpidites*. In contrast to Palynocomplex I, gymnosperm pollen from Palynocomplex II is enriched in such species as *Ephedrites* spp., *Sequoiapollenites pusillus* Singh, *Phyllocladites* sp., *Perinopollenites elatoides* Coup., *Exesipollenites* sp., and *Colliasporites trilobatus* (Balme) Dev.

The angiosperm pollen constitutes from 7 to 28% of all grains in palynospectra, and many of its species are quite representative (up to 5–7%). In addition to taxa occurring in Palynocomplex I, we also found here many forms typical of the upper Albian–Cenomanian of Europe, North America, and the Atlantic Ocean. These forms include *Perucipollenites minutus* Pacl., *Triporopollenites triangulus* Groot et Penny, *Striatopollis sarstedtensis* Kr., *Cretaceaiporites polygonalis* (Jard. et Magl.) Herngreen, *C. scabratus* Herngreen, *Tricolpoporopollenites* sp., *Nyssapollenites* sp., *Proaxoperites* sp., and *Tetracoplites* sp. In various combinations, they occur in the upper Albian–middle Cenomanian deposits of southern England and northern France (Laing, 1975), in the Albian and Cenomanian beds of Brazil (Herngreen, 1973), in the Albian–lower Cenomanian sediments of the northeastern Atlantic Ocean (Kotova, 1978), and in the middle–upper Albian strata of Canada (Singh, 1971).

Finally, *Psilatricolporites complanatius* Laing from Sample SK-9 was originally identified in the lower–middle Cenomanian deposits of France (the Sarta region; Laing, 1975).

BIOSTRATIGRAPHIC SYNTHESIS

Among the studied fossils occurring in the Paramonovo Clay, only foraminifers and spore–pollen assemblages provide the opportunity to refine the age of this stratigraphic unit.

By means of stratigraphic relations and data on agglutinated foraminifers and palynoflora, we may consider Member I as late Albian in age. Member II overlying it over the erosion surface bears *Hedbergella delrioensis* (Carsey). Regarding this taxon of planktonic foraminifers in a restricted volume (see above), we should state that its occurrence range is known beginning from the *Rotalipora ticinensis* Zone, which approximately corresponds to the terminal Albian, or Vraconian (the *Stoliczkaia dispar* Zone). Data on spores and pollen do not contradict this conclusion, because Palynocomplex I is typical of the upper Albian. A reliable dating of Member III is impossible, because it bears mostly agglutinated foraminifers, and rare specimens of coexisting calcareous forms are of a low stratigraphic value. However, Palynocomplex II, characterizing upper beds of this member, contains some Cenomanian and younger specimens (*Psilatricolporites complanatius* Laing and species of *Selaginella* genus). Therefore, it is not inconceivable that at least the upper part of Member III is of the Cenomanian age.

Unfortunately, the problem of the Albian–Cenomanian boundary is very complex, since the taxonomic changes in key stratigraphic fossils, such as ammonoids, planktonic foraminifers, and calcareous nannoplankton, are very gradual. Many taxa typical of the Cenomanian appeared already in the Vraconian. This implies that impoverished assemblages of the terminal Albian and lower Cenomanian, like those from the Paramonovo Formation, can be too similar for adequate discrimination.

LITHOLOGY, BIOFACIES, AND PALEOGEOGRAPHY

In the Spas-Kamenka section, the Paramonovo Formation is composed of silty clays with a minor proportion of fine-grained terrigenous material (Table 2). This sandy fraction is a the maximum content in Member III. In addition, nearly all of the studied samples contain fine (<0.1 mm) greenish grains of glauconite. The oxygonal shape of the grains suggests them to be clastic and derived from glauconite-rich sandy beds of the middle Albian. Nodular grains of syngenetic glauconite were distinguished only at the base of Member II, right above the hiatus. All samples, excluding SK-24, show the presence of white and greenish mica flakes. As it is known, the Paramonovo Clay is free of carbonates, and mostly consists of montmorillonite with minor zeolites (Kalitskaya, 1986). An elevated SiO_2 content in the clay (60–63.5% in the Spas-Kamenka section; Bizyaeva and Prints, 1974) is the result of an admixture of the aleuritic terrigenous quartz, opal radiolarian shells, and sponge spicules.

Silty–clayey sediments of this type are characteristic of cold-water shelves favorable for organisms with siliceous skeletons, but unpropitious for carbonate production.

As we noted above, Member III encloses thin (10 cm and less) lenticular interbeds of cross-bedded silt (Fig. 3). One, most persistent bed of this type at the base of the member (Sample SK-19) consists of a clay-free silty fraction enriched in fine quartz grains, glauconite, and mica. It does not contain fossils with the exception of fragmented sponge spicules. Light radiolarian skeletons seem to be washed out of this sediment under the influence of a strong current. The lithologic features of Member III are similar to those typical of sediments formed due to the storm activity in a basin of sedimentation, i.e., of tempestites. In particular, this is evident from the specific bedding textures of silty lenses. The storm events generate quite powerful, but irregular currents responsible for the deposition of coarse-grained sediments with a fine cross-lamination.

Tempestites occurring in Member III suggest the decreased depth of the past marine basin at the time of their accumulation. It is known that storm-generated currents do not penetrate deeper than 90–100 m. In vast epicontinental basins similar to the Central Russian Sea

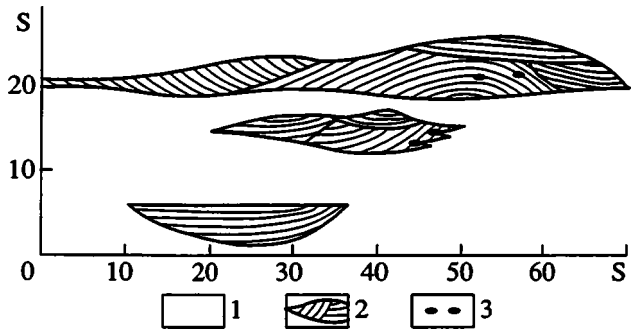

Fig. 3. Tempestites—lenticules of cross-bedded silts in the Paramonovo Clays:
1, clay; 2, silt; 3, rounded clay pebbles of gravel size.

of the Albian time, these currents should be active at a depth of about 40–50 m. Consequently, at the formation time of members I and II, the sea was deeper, probably as deep as 100 m.

The Paramonovo Clay contains authigenic pyrite in the form of rodlike and rounded, often globular microconcretions up to 1–2 mm in diameter. Their abundance is distinctly lower in the upper portion of Member III; however, it is unclear whether this is the result of original distribution, or a consequence of weathering and oxidation. Pyrite is an indicator of anaerobic conditions in sediments at the early diagenetic stage. The influence of reducing conditions on sediments is also evident from traces of the *Chondrites* life activity, which is indicative of such environments (Bromley and Ekdale, 1984); the traces are especially abundant in the lower member.

We failed to detect the macrofauna in the Paramonovo Formation exposed in the Spas-Kamenka section. These fossils seem to be lacking because of the oxygen deficiency in the bottom water of the basin, whose floor was mostly covered with unstable mud.

However, sporadically occurring sclecodonts (Sample SK-10), echinoid spines (Samples SK-17, SK-25, SK-27, and SK-28), prismatic and juvenile molluscan shells resembling *Inoceramus* and *Aucellina* (Samples SK-17 and SK-27), and fragmented shells of inarticulate lingulids (Sample SK-17) suggest that benthonic invertebrates were periodically able to dwell in the basin. Pyritized spicules of tetracrepid siliceous sponges are the most frequent organic remains; they are relatively rare in Member II and in the upper beds of Member III. These highly fragmented spicules seem to have been transported from tidal zones of the basin. Additional dwellers, which were mentioned above, were occasionally represented by ostracodes (Samples SK-17 and SK-27).

Agglutinated foraminifers were also significant components of benthic communities. Their taxonomically poor assemblages are distributed throughout the entire section of the Paramonovo Formation, sometimes in association with calcareous foraminifers. It is not very clear as to what extent the persistent occurrence of the latter in lower beds of Member II marks an episode of favorable conditions for these organisms; it could also be an effect of a higher preservation degree.

We have already mentioned that, in general, calcareous foraminifers are poorly preserved, partially dissolved, and recrystallized. Processes of corrosion also affected siliceous radiolarian skeletons. Therefore, it is not inconceivable that less abundant calcareous forms were completely dissolved in other parts of the section, either at the diagenetic stage of sediment evolution, or later, together with pyrite oxidation. This interpretation appears to be reasonable, because calcareous foraminifers sporadically occur at different levels of the section.

Radiolarians are the main planktonic fossils, and occur in all of studied samples with the exception of SK-19. In the grain fraction >0.1 mm, their abundance irregularly varies from 12 to 1397 specimen/g. It is at a maximum in Member I, where all estimated values are above 450 specimen/g, including the peak value of 1397 specimen/g calculated for Sample SK-33. The abundance of radiolarians in Member II changes from 61 to 819 specimen/g, despite the perfect preservation of fossils in this interval. Minimum values (18–83 specimen/g) were identified in the lower part of Member III (samples SK-11 to SK-17); higher, the abundance of radiolarians grows up to 467 specimen/g at the top (Sample SK-2).

The taxonomic composition and morphology of radiolarians indicate that the basin of their habitat was of the neritic type. At the same time, it should be deep enough and open to maintain desirable conditions for the existence of radiolarians.

Planktonic foraminifers were distinguished at only one level (Sample SK-27), where they constitute a few percent of the assemblage. Their presence also points to the considerable depth of the Central Russian Sea that formerly existed near Moscow. The lower Oxfordian is another stratigraphic interval in this region marked by the occurrence of planktonic radiolarians.

It is very likely that diatoms were also significant components of marine biota, because their pyritized casts are dispersed throughout the studied section. In Member I and throughout a greater part of Member II, we identified large cysts of pelagic algae *Leiosphaeridia*, which are typical of sediments deposited in anoxic environments (Leckie *et al.*, 1992). Concurrently, our study of spore–pollen samples showed that dinoflagellate cysts are scarce.

The sea was the habitat of fishes, whose scales and bones, though not abundant, occur within members I and II. Finally, the Paramonovo Clay contains a notable amount of coalified plant detritus, which was derived from bordering land areas.

Summarizing all the data, we may conclude that the Central Russian Sea of the late Albian time was of normal salinity and considerably deep (40–100 m), though shoaling with time. The oxygen concentration near the

Table 2. Distribution of microfossils and some mineral components through the Paramonovo Formation, the Spas-Kamenka section

| Components | Sample numbers | | |

bottom was low, and this negatively affected the benthic community.

Some aspects of the paleogeography at the accumulation time of the Paramonovo Formation and the general evolution of the Russian platform were discussed by Sazonova (1958), Derviz (1959), Gerasimov *et al.* (1962), Sazonova and Sazonov (1967). The late Albian paleogeography was of special interest for Baraboshkin (1991), though many of his results are still unpublished.

Formerly, the total area of the Russian platform, which was covered with clay deposits bearing radiolarians, was very large; however, the subsequent erosion left only isolated patches of these sediments. The largest preserved area comprising type sections of the Paramonovo Formation is to the east of Moscow. Pertinent deposits are traceable eastward and northeastward from here, in the Yaroslavl' and Vladimir regions (Gerasimov, 1971, 1988; Olfer'ev, 1986). In the Ryazan–Saratov depression, the upper Albian clays with radiolarians are exposed in the Putyatin region (Sazonova, 1958; Sazonova and Sazonov, 1967). A sequence of clays identical in their lithology and radiolarian assemblage to the Paramonovo Formation is widespread in the Tambov and Penza regions. It extends southeastward of this area into the Ul'yanovsk, Saratov, and Samara regions (Sazonova and Sazonov, 1967; Glazunova, 1988). In Baraboshkin's samples from the Vol'sk and Uvek localities (the last one near the town of Saratov), we identified an assemblage of radiolarians and agglutinated foraminifers, which are indistinguishable from the Paramonovo assemblage. It is also likely that a corresponding basin of sedimentation occupied a considerable area in the peri-Caspian depression. Dark clays with radiolarians and/or agglutinated foraminifers are known near Novouzensk, in the Inder–Chelkar region, and in the Sagiz–Emba and southern Emba river basins (Sazonova and Sazonov, 1967; Koltypin *et al.*, 1988; Romashov and Chertova, 1972).

In addition, isolated areas with the upper Albian clays often bearing radiolarians are exposed westward of Moscow. They were first described in detail near Bryansk (Bushinskii, 1954), and presumably occur in the Kaluga and Smolensk regions, although corresponding rocks are not very well studied here. Farther to the west, in the Belarussian territory, the upper Albian beds consist of silts and shallow-water sands (Akimets, 1974), which are free of foraminifers, but locally bear *Aucellina gryphaeoides* (Sow.). However, at present, there are no reliable data to suggest that true analogues of the Paramonovo Formation really exist in Belarus.

The Central Russian sea of the late Albian time was definitely connected with the Transcaspian basin in the southeast, and with the Crimea-Caucasus basin in the south. It is also conceivable that in the west, it was adjoining with the Poland–Lithuania Sound of the Cretaceous sea.

The paleogeography of the northern East European platform is insufficiently clear, because respective deposits were eliminated here by the later erosion. However, recently, the upper Albian–Cenomanian clays with the foraminifer species *Verneuilinoides perlexus* and abundant radiolarians were distinguished in the Barents Sea (Vassilenko, 1992).

DEPTH FLUCTUATIONS AND EUSTATIC RHYTHM IN THE MID-ALBIAN BASINS OF THE NORTHERN MOSCOW REGION

As was shown above, the Paramonovo Clay northward of Moscow was accumulated in an epicontinental marine basin, which was as deep as 90–100 m. The immense distribution area of these deposits in the Central Russian platform and near the Caspian Sea discredits the idea that the origin of the basin was related to the local subsidence of the earth's crust. Therefore, the terminal late Albian transgression in the Russian platform was of eustatic origin. To understand the importance of this eustatic event and the corresponding sea-level rise, we should compare it with earlier and later eustatic fluctuations.

As far as it concerns the present study, we restrict the comparative analysis to the Aptian–Cenomanian interval (Fig. 4). Unfortunately, the precise dating of some events is impossible, because many deposits of this interval are barren of fossils with definite biostratigraphic significance.

We believe that in the northern Moscow region, the 30-m-thick Aptian sequence consists of sandy alluvial (in the lower part) and presumably deltaic (in the upper part) deposits, although Olfer'ev (1988) suggested all of them to be of a shallow-marine genesis. These deposits are subdivided into the Iksha, Vorokhobino, and Volgusha formations (Olfer'ev, 1986). In the opinion of Olfer'ev (1986), the range of the two lower formations corresponds to an upper part of the lower Aptian. The Iksha Formation encloses organic remains well known from publications as the "Klin," "Tatarovo," or "Karovo" flora. However, it is impossible to precisely date this flora. According to data on the western Vladimir region (Olfer'ev, 1986), the Vorokhobino Formation yielded here the early Aptian ammonite *Aconeceras trautscholdi* (Sinzov). The Volgusha Formation overlying this unit over the erosion surface is referred to the upper Aptian on the basis of spore–pollen data. Despite the very impoverished characteristics of this palynoflora, its position between the Barremian and Albian is beyond a doubt, and the accumulation time of host deposits corresponds to the commencement of a transgressive cycle, when a higher sea level initially caused flooding in overdeepened river valleys, and then migration of coastal zones of the expanding sea inside the continent. In the northern Moscow region, the topmost Aptian beds are exposed at the absolute height approaching +100 m. If marine waters penetrated into this area from the Volga region,

which represented an open sea basin of that time, the basin depth in the Moscow region should be at the minimum, and the height of +110 m should correspond to the peak value of the Aptian sea level. Taking into account the depth of the flooded river valleys, we have to admit that the initial height of the sea level was at about +80 m.

The northern Moscow region lacks the lower Albian deposits, as they were eroded during the middle Albian transgression. Nevertheless, their traces are perceptible in redeposited phosphorites occurring at the base of the Gavrilkovo Formation and bearing ammonites of the lower Albian biozones (Baraboshkin, 1991, 1992). Eastward, in the Vladimir region and Ryazan'–Saratov depression (Penza region), shallow-marine sands with lower Albian phosphorites are as thick as 15–20 m. This means that a shallow sea of that time with water level at a point between +125 and +130 m, which is necessary to admit the above thickness of accumulated sediments, really existed northward of Moscow.

After a temporary, but sharp middle Albian fall of the sea level, the latter raised again up to +150 or +160 m. This considerably enlarged the area of the basin, which still represented a shallow sea with a depth range of 10–20 m. This stage is marked by the accumulation of the Gavrilkovo Formation, consisting of glauconite sands with several horizons of phosphorite concretions. The concretions were likely deposited during the moments of basin shoaling and reworking of the older sediments, and the formation can be accordingly referred to the condensed type. The top of the Gavrilkovo deposits is near the point of +140 m, and therefore, the respective sea level was at least approaching +150 m.

The first half of the late Albian presumably was a time corresponding to the break in sedimentation. After this, vast areas of the Russian platform were covered by the relatively deep (40–100 m) and cold marine basin, which accumulated clays rich in radiolarians. The sea level during this eustatic event should be approaching +240 m, i.e., 80–90 m higher than its middle Albian position. We estimated these values using the average hypsometric mark of the Paramonovo Formation base (+140 m), and adding to it the inferred initial depth of the basin (90–100 m). The lesser sea depth (40–50 m) during the accumulation period of the upper half of the Paramonovo Formation was a possible response to either the eustatic sea-level fall, or the fast accumulation of clayey sediments. It is unclear which of these factors was the most important, and of course, their summary effect could be responsible for shoaling. Near the village of Bogorodskoe (northeastward of Sergiev Posad), the top of the Paramonovo Formation lies between +175 and +177 m. The difference between these values and the sea-level peak (+240 m), i.e., approximately 60 m, corresponds to the assumed sea-level decrease.

In the northern Moscow region, the lower–middle Cenomanian deposits are represented by the Lyaminskii Formation. It consists of coarse- and medium-grained sands enclosing the phosphorite horizon at the base, which yielded ammonites (*Schloenbachia*), bivalve mollusks, inarticulate brachiopods, and shark teeth (Milanovskii, 1926). The formation rests on the erosion surface of the underlying sands of the Yakhroma Formation closely interrelated with the Paramonovo Formation. Though generally accepted, the Cenomanian age of the Lyaminskii Formation has never been well argued, because its phosphorite concretions with Cenomanian fossils could be of the redeposited type, which is typical of other Cretaceous deposits in central Russia (Baraboshkin, 1991). Consequently, the host sands can be younger, for instance, Turonian in age.

Disregarding this age problem, we still may state that a shallow Cenomanian sea was spread over the northern Moscow region, as it is shown in Fig. 4. During this transgressive phase, the eustatic rise of sea level was approaching the absolute height of +200 m.

Despite some indications (Dobrov, 1932; Dan'shin, 1947), the Turonian deposits are missing in the Moscow region. The Khot'kovo Opoka rich in radiolarians is of the late Santonian–early Campanian age (Bragina, 1994). According to Olfer'ev (1986), the upper Turonian–Coniacian interval corresponds to chalklike marls bearing foraminifers and occurring near the Pereyaslavl'-Zalesskii, the Yaroslavl'–Vladimir region. However, the poorly described microfossils from these rocks need to be revised.

Thus, the terminal Albian transgression was maximal with respect to the area covered by the sea and the basin depth. This was the termination of a transgressive cycle initiated in the Aptian Age. In the Russian platform, comparable events are detectable only in the middle–upper Oxfordian, middle Volgian, and upper Santonian–lower Campanian deposits.

The described transgression of the terminal Albian time is correlative with similar events in other regions. For instance, the middle–late Albian transgressive episodes were also recognized in some regions of the East European platform and its southern margin (Naidin *et al.*, 1980).

It is important to understand the reliability of the estimated values exemplifying the eustatic sea-level fluctuations during the mid-Cretaceous time. First of all, they are based on the assumption that the northern Moscow region was tectonically stable at that time, and vertical movements of the earth's crust represented only isostatic reactions to either the accumulation of sediments, or the removal of sea water during the regressive phases. However, the idea on the extreme stability of the Russian platform (Sahagian and Holland, 1991) is incompatible with the available data, and has already been subjected to critical comments concerning the Cenomanian–Turonian boundary deposits (Blank *et al.*, 1992). The facial analysis of the Lower Cretaceous deposits throughout the Russian platform

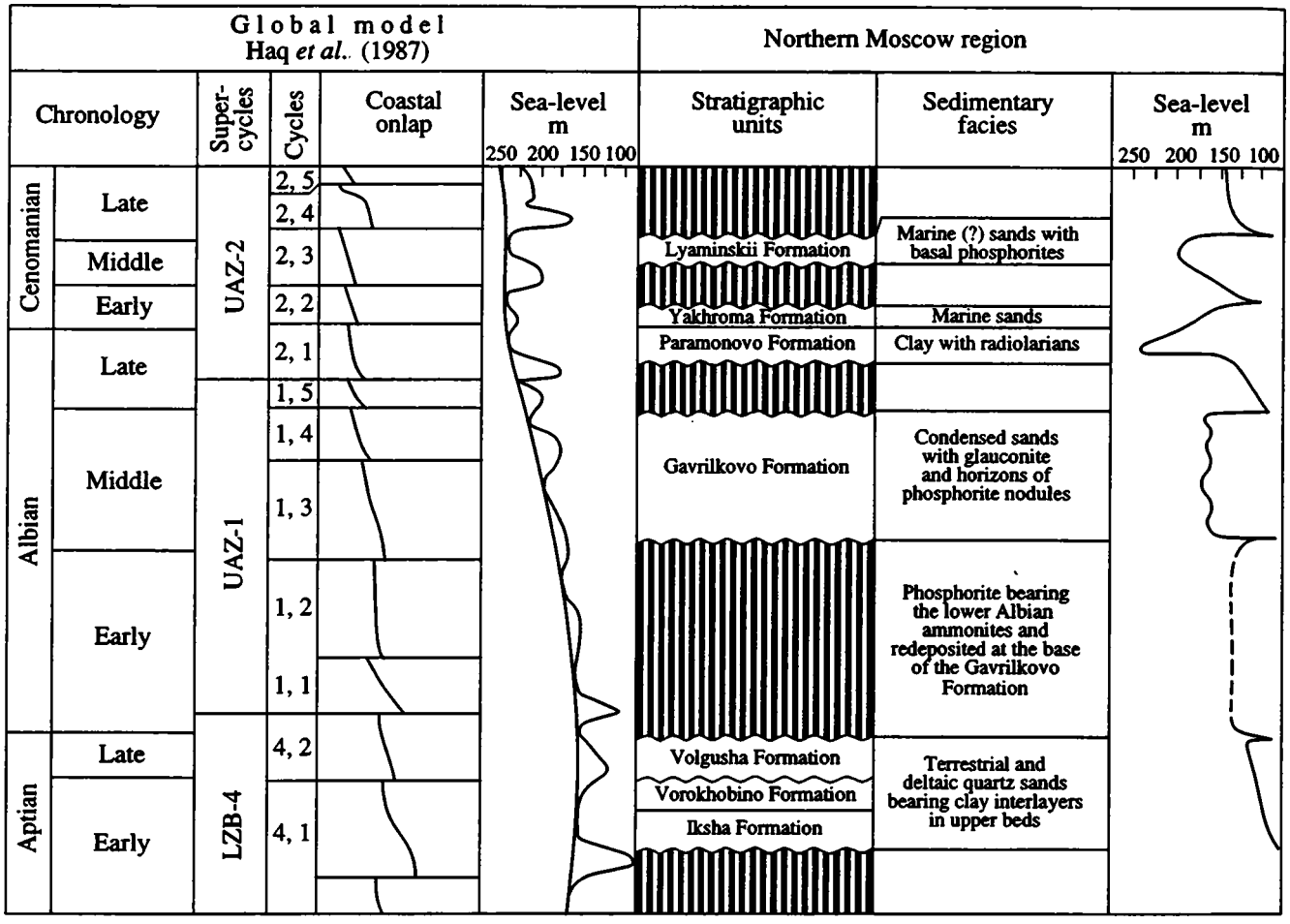

Fig. 4. Aptian–Cenomanian eustatic fluctuations of sea level in the northern Moscow region.

(Sazonova and Sazonov, 1967) also showed that many structural elements, such as the Moscow syneclise, Ryazan –Saratov depression, Oka–Tsna rampart, and others were still tectonically active at that time. They experienced opposite vertical movements, as is evident from the respective changes in the positions of different facial zones. The raise and subsidence amplitudes are unknown. As far as this concerns the Moscow region, we may note that the Upper Jurassic deposits lie considerably lower (for 50–60 m) in the Klin–Dmitrov ridges, than in the Teplyi Stan Hills (the southwestern district of Moscow). At the same time, the upper Volgian and Berriasian deposits are lacking everywhere in the north, and this indicates an uplift of the northern Moscow region during the Neocomian.

Despite these data, Olfer'ev (oral presentation in 1994) is of the opinion that all of the aforementioned structures were absolutely passive in the Early Cretaceous period.

Differentiated block movements were nevertheless characteristic of the very recent, neotectonic period (Gavryushova *et al.*, 1986). Their amplitudes were up to few, or even 10 m.

Discrediting the idea on the tectonic stability of the central Russian platform, the above data show that the reconstructed curve of sea-level fluctuations is not very valid. Of course, we may assume that opposite movements during certain periods of time are insignificant for our calculations. Nevertheless, it is also not inconceivable that the obtained values include a systematic error of unknown sign. To reveal the amplitudes and sign of tectonic dislocations, one has to analyze and compare the well-dated sedimentary sections of different blocks.

At present, we set aside the problem of tectonic movements in the northern Moscow region, and accept the calculated values, as if they were realistic. The standard curve of eustatic sea-level fluctuations by Haq *et al.* (1987, 1988) is now popular among scientists. Comparing it with our curve depicted for the northern Moscow region (Fig. 4), we may see that both are similar in general features. According to data by Haq *et al.*, the terminal Albian transgression was the real event, and the respective absolute height of the sea level was near +250 m, thus very close to the value estimated for the Moscow region (+240 m). However, one can see

that the reference curve of Haq *et al.* suggests nearly the same amplitudes of global sea-level rises in the early and middle Cenomanian time, and even a higher level for the early Turonian event. If the available age determinations for mid-Cretaceous formations of the northern Moscow region are reliable, then we may conclude that the early Turonian event did not affect this area. The maximum sea-level rise is confined here to the terminal Albian, and perhaps, the earliest Cenomanian, if we correctly consider the last interval as corresponding to the upper part of the Paramonovo Formation.

Analyzing the Klimovka, Inza, and Lomov sections in the Volga region, Sahagian and Holland (1991) concluded that the sea level rose a total of 120 m during the Bajocian–Cenomanian interval. Their graph demonstrates a continuous sea-level rise throughout the Early Cretaceous (for 100 m in total); the rise is interrupted only by the early Albian fall (at least for 17.5 m).

There is a disagreement between these results and data on the northern Moscow region. This territory did not recorded the continuous rise of the sea level. The lower Albian deposits mark here the further progress of the transgression initiated in the Aptian time. For the period from the middle to terminal Albian, the inferred sea-level rise was 80–90 m. Consequently, the sea-level fluctuations were considerably more frequent and intense than was formerly believed.

CONCLUSION

The terminal Albian transgression was very widespread in the central regions of the Russian platform. It promoted the migration of planktonic foraminifers in this area, was favorable for the intense evolution of radiolarians, and can be termed the *Paramonovo event*. According to some data, this event also appears to be traceable in northern Europe. For instance, despite the indication of Hancock (1990) that in the British Islands, the late Albian sea level was relatively low and stable, micropaleontological studies demonstrated the abundance increase of planktonic forms among foraminifers during the late Albian time that was especially pronounced in the *Stoliczkaia dispar* Zone (Hart, 1973). Other data showed that the Spiculite Member rich in radiolarians is regularly associated with the terminal Albian–lower Cenomanian marls exposed in several well-studied sections of northern Germany (Frieg and Kemper, 1989). Despite the fact that this member shows variable stratigraphic ranges in different localities, its formation seems to be undoubtedly related to the Paramonovo event.

Available data also suggest that the terminal Albian radiolarians may occur in such European regions as northern Bulgaria and the northern Caucasus; however, this assumption has yet to be adequately proved.

Obviously, one can evaluate the real significance of the Paramonovo event only by means of additional detailed regional data.

Reviewer I.A. Basov

REFERENCES

Alexandrowicz, S.W., Otwornice albu serii czorsztynskiej (formacja z Chmielowej) Pieninskiego pasa skalkowego, *Roczn. Polsk. Towarz. Geol.*, 1979, vol. 49, nos. 1–2, pp. 165–183.

Akimets, V.S., Albian Deposits of Belarus', in *Nekotorye voprosy stratigrafii i paleontologii paleozoiskikh i mezozoiskikh otlozhenii Belorussii* (Some Problems of Stratigraphy and Paleontology of Paleozoic–Mesozoic Deposits in Byelorussia), Minsk: Bel. Nauch.-Issled. Geol.-Razv. Inst., 1974, pp. 233–249.

Aliev, Kh.Sh. and Smirnova, R.F., New Radiolarian Species from the Albian Deposits of the Central Russian Platform, in *Iskopaemye i sovremennye radiolyarii* (Fossil and Recent Radiolarians), Lvov: Lvov Univ., 1969, pp. 62–72.

Baraboshkin, E.Yu., Albian Stratigraphy and Ammonites of the Russian Plate, *Cand. Sci. (Geol.–Mineral.) Dissertation*, Moscow: Moscow State Univ., 1991, p. 19.

Baraboshkin, E.Yu., Lower Albian in Central Areas of the Russian Plate, in *Stratigrafiya fanerozoya tsentra Vostocho-Evropeiskoi platformy* (Phanerozoic Stratigraphy of Central East European Platform), Moscow: Tsentrgeologiya, 1992, pp. 20–36.

Baraboshkin, E.Yu. and Mikhailova, I.A., Ammonites and Middle Albian Stratigraphy of the Northern Moscow Region. Part 1: Stratigraphy, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1987, vol. 62, no. 6, pp. 92–100.

Bizyaeva, O.A. and Prints, P.N., Clay and Loam, in *Geologiya SSSR, T. 4: Tsentr evropeiskoi chasti SSSR, Poleznye iskopaemye* (Geology of the USSR, vol. 4: Mineral Resources in the Central European Part of the USSR), Moscow: Nedra, 1974, pp. 148–180.

Blank, M.Ya., Naidin, D.P., and Olfer'ev, A.G., The Cenomanian Roof Relief in the Dnieper–Donets Depression, Voronezh Anteclise, and Adjacent Structures of the East European Platform, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1992, vol. 67, no. 6, pp. 43–47.

Bragina, L.G., Radiolarians and Stratigraphy of the Upper Cretaceous Deposits, the Khot'kovo Group of the Moscow Region, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1994, vol. 69, no. 2, pp. 91–100.

Bromley, R.G. and Ekdale, A.A., Chondrites: A Trace Fossil Indicator of Anoxia in Sediments, *Science*, 1984, vol. 224, pp. 872–874.

Brotzen, F., De geologiska resultaten frün borrningarna vid Höllviken. Preliminär rapport. Del. 1. Kritan, *Sver. Geol. Undersökn., Ser. C*, 1945, no. 465, pp. 1–65.

Bukalova, G.V., Anomalinids from Albian Deposits of the Laba–Urukh Interstream Area (Northern Ciscaucasia), *Tr. Vses. Nauch.-Issled. Geol.-Razv. Inst.*, 1958, vol. 9, pp. 177–196.

Bushinskii, G.I., Lithology of Cretaceous Deposits in the Dnieper–Donets Depression, *Tr. Inst. Geol. Nauk*, 1954, issue 156.

Caron, M., Cretaceous Planktic Foraminifera, in *Plankton Stratigraphy*, Cambridge: Cambridge Univ., 1985, pp. 17–86.

Couper, R.A., Upper Mesozoic and Cenozoic Spores and Pollen Grains from New Zealand, *Paleontol. Bull. N.Z. Geol. Surv.*, 1953, no. 22.

Cushman, J.A. and Alexander, C.I., Some Vaginulinas and Other Foraminifera from the Lower Cretaceous of Texas, *Contrib. Cushm. Lab. Foram. Res.*, 1930, vol. 6, pp. 1–10.

Dan'shin, B.M., *Geologicheskoe stroenie i poleznye iskopaemye Moskvy i ee okrestnostei* (Geological Structure and Mineral Resources of Moscow and Surrounding Areas), Moscow: Mosk. O-vo Ispyt. Prir., 1947.

Derviz, T.L., Volga–Urals Oil-Bearing Region: Jurassic and Cretaceous Deposits, *Tr. Vses. Nauch.-Issled. Geol.-Razv. Inst.*, 1959, issue 145.

Dobrov, S.A., Geology of the Dmitrov Region, *Tr. Mus. Dmitrov. Kraya*, 1932, issue 7, pp. 5–131.

Frieg, C. and Kemper, E., Micropalaontologische Gliederung und Abgrenzung von Ober-Alb und Unter-Cenoman in Nordwestdeutschland, *Geol. Jahrb.*, 1989, vol. A113, pp. 73–193.

Frizzell, D.L., Handbook of Cretaceous Foraminifera of Texas, *Publ. Univ. Tex., Bur. Econ. Geol.*, 1954, no. 22.

Gavryushova, E.A., Lekht, E.E., and Fedonkina, I.N., Linear Neotectonic Structures of the Moscow Region (Results of Satellite Image Deciphering), in *Geologiya i poleznye iskopaemye tsentral'nykh raionov Vostochno-Evropeiskoi platformy* (Geology and Mineral Resources of the Central East European Platform), Moscow: Mosk. O-vo Ispyt. Prir., 1986, pp. 110–117.

Gerasimov, P.A., Cretaceous System, Lower Series, in *Geologiya SSSR, T. 4: Tsentr evropeiskoi chasti SSSR, Chast' 1: Geologicheskoe opisanie* (Geology of the USSR, vol. 4: Central European Part of the USSR, Part 1: Geological Description), Moscow: Nedra, 1971, pp. 425–445.

Gerasimov, P.A., Cretaceous System, Lower Series: Southern Parts of the Moscow Syneclise and Voronezh Anteclise, in *Stratigrafiya SSSR, Melovaya sistema, Polutom 1* (Stratigraphy of the USSR, Cretaceous System, Semivolume 1), Moscow: Nedra, 1988, pp. 55–60.

Gerasimov, P.A., Migacheva, E.N., Naidin, D.P., and Sterlin, B.P., *Yurskie i melovye otlozheniya Russkoi platformy* (Jurassic and Cretaceous Deposits of the Russian Platform), Moscow: Mosk. Gos. Univ., 1962.

Glazunova, A.E. and Poslavskaya, G.G., Cretaceous System, Lower Series: Ryazan–Saratov Depression and Volga–Urals Anteclise, in *Stratigrafiya SSSR, Melovaya sistema, Polutom 1* (Stratigraphy of the USSR, Cretaceous System, Semivolume 1), Moscow: Nedra, 1988, pp. 61–66.

Gorbachik, T.N., Planktonic Foraminifers from Lower Cretaceous Deposits of the North Atlantic, in *Stratigrafiya i paleogeografiya Severnoi Atlantiki v melovom periode* (Stratigraphy and Paleogeography of North Atlantic during the Cretaceous Period), Leningrad: Nauch.-Issled. Inst. Geol. Arktiki, 1980, pp. 29–45.

Hancock, J.M., Sea-level Changes in the British Region during the Late Cretaceous, *Proc. Geol. Assoc.*, 1989 (1990), vol. 100, no. 4, pp. 565–594.

Haq, B.U., Hardenbol, J., and Vail, P.R., The Chronology of the Fluctuating Sea Level since the Triassic, *Science*, 1987, vol. 235, pp. 1156–1167.

Haq, B.U., Hardenbol, J., and Vail, P.R., Mesozoic and Cenozoic Chronostratigraphy and Cycles of Sea-Level Change, *Sea-Level Changes: An Integrated Approach*, Wilgus, C.K., Hastings, B.S., Kendal, C.G.St.C., *et al.*, Eds., Spec. Publ. Soc. Econ. Paleontol. Mineral., 1988, no. 42, pp. 71–108.

Hart, M.B., A Correlation of the Macrofaunal and Microfaunal Zonations of the Gault Clay in Southeast England, *The Boreal Lower Cretaceous*, Casey, R. and Rawson, P.F., Eds., Liverpool: Seel House, 1973, pp. 267–288.

Herngreen, G.F.W., Palynology of Albian–Cenomanian Strata of Borehole 1-QS-1-MA, State of Maranho, Brazil, *Pollen Spores*, 1973, vol. 15, nos. 3–4, pp. 515–555.

Juhasz, M., Palynostratigraphic Zonation of the Transdanubian Middle Cretaceous, *Acta Geol. Hung.*, 1983, vol. 26, nos. 1–2, pp. 41–68.

Kazintsova, L.I., Radiolarians, in *Opornyi razrez melovykh otlozhenii Sakhalina (Naibinskii razrez)* [Reference Section of Cretaceous Deposits in Sakhalin (the Naiba Section)], Leningrad: Nauka, 1987, pp. 93–97.

Kalutskaya, S.A., A perspective in Using Some Nonmetalliferous Deposits in Agriculture, in *Geologiya i poleznye iskopaemye tsentral'nykh raionov Vostochno-Evropeiskoi platformy* (Geology and Mineral Resources of the Central East European Platform), Moscow: Mosk. O-vo Ispyt. Prir., 1986, pp. 145–152.

Khimenkov, V.G., Problems of Studying Ground Waters of Moscow, in *Artezianskie vody g. Moskvy: Snabzhenie imi ee naseleniya i promyshlennykh predpriyatii* (Artesian Waters of Moscow: Supply for Population and Industry), Moscow: Moskomkhoz, 1928, pp. 7–16.

Koltypin, S.N., Mjatliuk, E.V., and Poslavskaya, G.G., Cretaceous System, Lower Series: Caspian Syneclise, in *Stratigrafiya SSSR, Melovaya sistema, Polutom 1* (Stratigraphy of the USSR, Cretaceous System, Semivolume 1), Moscow: Nedra, 1988, pp. 66–74.

Kotova, I.Z., Spores and Pollen from Cretaceous Deposits of the Eastern North Atlantic Ocean, Deep Sea Drilling Project Leg 41, Sites 367 and 370, *Initial Rep. Deep Sea Drill. Project*, Washington: US Gov. Print. Off., 1978, vol. 41, pp. 841–881.

Krasnova, L.N., Selaginellaceae, in *Pyl'tsa i spory Zapadnoi Sibiri: Yura–Paleotsen* (Pollen and Spores from West Siberia: Jurassic–Paleocene), Moscow: Nedra, 1961, pp. 19–44.

Krasheninnikov, V.A. and Basov, I.A., *Stratigrafiya mela Yuzhnogo okeana* (Cretaceous Stratigraphy of the Southern Ocean), Moscow: Nauka, 1985.

Laing, J.F., Mid Cretaceous Angiosperm Pollen from Southern England and Northern France, *Palaeontology*, 1975, vol. 18, no. 4, pp. 775–808.

Leckie, D.A., Singh, C., Bloch, J., *et al.*, An Anoxic Event at the Albian–Cenomanian Boundary: the Fish Scale Marker Bed, Northern Alberta, Canada, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 1992, vol. 92, nos. 1–2, pp. 139–166.

Loeblich, A.R. and Tappan, H., Foraminifera from the Walnut Formation (Lower Cretaceous) of Northern Texas and Southern Oklahoma, *J. Paleontol.*, 1949, vol. 23, no. 3, pp. 245–266.

Magniez-Jannin, F., *Les Foraminiferes de l'Albien de l'Aube: Paleontologie, Stratigraphie, Ecologie*, Paris, 1975.

Milanovskii, E.V., The Cenomanian Deposits of the Moscow Province, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1926, vol. 4, nos. 1–2, pp. 75–79.

Mjatliuk, E.V. and Vassilenko, V.P., *Atlas kharakternykh foraminifer nizhnemelovykh otlozhenii Prikaspiiskoi nizmen-*

nosti, p-va Mangyshlak i Ustyurta (Atlas of Characteristic Lower Cretaceous Foraminifers from the Peri-Caspian Lowland, Mangyshlak Peninsula, and Ustyurt), Leningrad: Nauka, 1988.

Moullade, M., Etude Stratigraphique et Mocropaleontologique du Cretace Inferieur de la "Fosse Vocontienne," *Doc. Lab. Geol. Fac. Sci. Lyon*, 1966, no. 15.

Naidin, D.P., Sazonova, I.G., Poyarkova, Z.N., *et al.*, Cretaceous Transgressions in the East European Platform, Crimea, and Central Asia, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1980, vol. 55, no. 5, pp. 27–42.

Olfer'ev, A.G., New Data on Geological Structure of the Lower Cretaceous Deposits in Moscow Region, in *Geologiya i poleznye iskopaemye tsentral'nykh raionov Vostochno-Evropeiskoi platformy* (Geology and Mineral Resources of the Central East European Platform), Moscow: Mosk. O-vo Ispyt. Prir., 1986, pp. 44–55.

Olfer'ev, A.G., Structure of Lower Cretaceous Deposits in Central and Southern Areas of Moscow Syneclise and Distribution of Mineral Resources, *Cand. Sci. (Geol.–Mineral.) Dissertation*, Moscow: Mosk. Geol.-Razved. Inst., 1988.

Pačltova, B., Cretaceous Angiosperms of Bohemia, *Cent. Europ. Bot. Rev.*, 1977, vol. 43, no. 1, pp. 128–192.

Riegraf, W. and Luterbacher, H.-P., Bentonische Foraminiferen aus der Unterkreide des "Deep Sea Drilling Project" (Legs 1–79), *Geol. Rundsch.*, 1989, vol. 78, no. 3, pp. 1063–1120.

Risch, H., Stratigraphie der hoheren Unterkreide der Bayerischen Kalkalpen mit hilfe von Microfossilen, *Palaeontographica*, 1971, vol. A138, nos. 1–4, pp. 1–80.

Rozanov, A.N., Stratigraphy of Phosphorite-bearing Deposits in Central and Northeastern Regions, Phosphorite Horizons, and Their Viable Origin, in *Fosfority SSSR* (Phosphorites of the USSR), Leningrad: Geolkom, 1927, pp. 133–150.

Romashov, A.A. and Chertova, S.O., Marine Upper Albian Deposits in the Eastern Peri-Caspian Depression, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1972, vol. 47, no. 6, pp. 38–44.

Sahagian, D.L. and Holland, S.M., Eustatic Sea-Level Curve Based on a Stable Frame of Reference: Preliminary Results, *Geology*, 1991, vol. 19, no. 12, pp. 1209–1212.

Sahagian, D.L. and Jones, M., Quantified Middle Jurassic to Paleocene Eustatic Variations Based on Russian Platform Stratigraphy: Stage Level Resolution, *Bull. Geol. Soc. Am.*, 1993, vol. 105, no. 8, pp. 1109–1118.

Sazonova, I.G., Lower Cretaceous Deposits in Central Areas of the Russian Platform, in *Mezozoiskie i tretichnye otlozheniya tsentral'nykh oblastei Russkoi platformy* (Mesozoic and Tertiary Deposits in Central Areas of the Russian Platform), Moscow: Gostoptekhizdat, 1958, pp. 31–184.

Sazonova, I.G. and Sazonov, N.T., Paleogeography of the Russian Platform during the Jurassic and Early Cretaceous Time, *Tr. Nauchno-Issled. Inst. Geol. Nefti*, 1967, issue 62.

Sigal, J., Contribution a une monographie des rosalines. 1. Le genre *Ticinella* Reichel, souche de rotalipores, *Eclogae Geol. Helv.*, 1966, vol. 59, no. 1, pp. 185–218.

Singh, C., Lower Cretaceous Microfloras of the Peace River Area, Northwestern Alberta, *Bull. Res. Counc. Alberta*, 1971, vol. 1, no. 28.

Smirnova, S.B. and Aliev, O.B., Spore–Pollen Complexes from the Cenomanian and Coniacian Deposits of the Lesser Caucasus, in *Problemy paleontologii i stratigrafii Azerbaidzhana* (Problems of Paleontology and Stratigraphy of Azerbaijan), Baku: Elm, 1984, pp. 40–48.

Sokolov, V.D., Organization and General Progress in Studies within the Moscow District with Respect to Hydrogeology and Better Water Supply for Settlements, in *Gidrogeologicheskii ocherk Moskovskoi gubernii* (Hydrogeology of the Moscow District), Moscow: Pechatnya S.P. Yakovleva, 1913, pp. 3–19.

Tappan, H., Foraminifera from the Grayson Formation of Northern Texas, *J. Paleontol.*, 1940, vol. 14, no. 2, pp. 93–126.

Tappan, H., Foraminifera from the Duch Creek Formation of Oklahoma and Texas, *J. Paleontol.*, 1943, vol. 17, no. 4, pp. 476–517.

Tappan, H., New Cretaceous Index Foraminifera from Northern Alaska, *Bull. US Nat. Mus.*, 1957, no. 215, pp. 201–222.

Ten Dam, A., Les foraminiferes de l'Albien des Pays-Bas, *Mem. Soc. Geol. France*, 1950, no. 63.

Vail., P.R., Mitchum, R.M., and Thompson, S., Seismic Stratigraphy and Global Changes of Sea Level, Part 4: Global Cycles of Relative Changes of Sea Level, *Seismic Stratigraphy—Applications to Hydrocarbon Exploration*, Payton, C.E., Ed., *Mem. Am. Assoc. Petrol. Geol.*, 1977, no. 26, pp. 83–97.

Vassilenko, V.P., Upper Cretaceous Foraminifers from the Mangyshlak Peninsula, *Tr. Vses. Nauch.-Issled. Geol.-Razv. Inst.*, 1961, vol. 171.

Vassilenko, L.V., Comparative Characteristics of Albian Foraminifer Fauna from the Barents and Kara Shelves, in *Geologicheskaya istoriya Arktiki v mezozoe i kainozoe, Kniga 1* (Mesozoic–Cenozoic Geological History of Arctic Regions, Book 1), St. Petersburg: VNIIOkeangeologiya, 1992, pp. 71–77.

Vishnevskaya, V.S., Late Mesozoic Radiolarian Beds of the USSR, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1991, no. 2, pp. 58–81.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 362–370. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 53–61.
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Paleogene Radiolarians of Southern Russia: Composition and Paleoecological Interpretations

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Received April 4, 1994; in final form, December 20, 1994

Abstract—Radiolarians from seven Paleogene sequences in different structural-facial zones of the northern Tethyan epicontinental areas were investigated. Seven radiolaria-bearing intervals, well-traceable through the studied region, were defined. These intervals mark the well-known Paleogene transgressions. Analysis at the family and subfamily levels was performed for the middle and late Lutetian radiolarian assemblages from the northern Caspian and eastern Ciscaucasia regions. Based on the results, paleogeographic interpretations were suggested.

Key words: *Paleogene, stratigraphy, radiolarians, taxonomic composition, paleoecology, southern Russia, northern Caspian region.*

Radiolarians from the Paleogene deposits of the northern and eastern Tethys were discovered in the early 1950s (Lipman, 1950, 1953). Later, many researchers investigated their taxonomic composition and stratigraphic distribution (Borisenko, 1960a, 1960b; Krasheninnikov, 1960; Zagorodnyuk, 1970; Gorbunov, 1979; Tochilina, 1975, and others). As a result, several local stratigraphic schemes based on radiolarians were suggested to subdivide the Paleogene deposits in different areas of the Russian platform, Ciscaucasia, and Middle Asia. In 1989, the Interdepartmental Stratigraphic Commission (*Resheniya* ..., 1989) adopted the stratigraphic subdivisions of the Paleocene–Eocene deposits in the Russian platform, which were based on the zonal scale by Kozlova (1984, 1990).

Paleogene radiolarians occur in many regions of the former Soviet Union, representing the southern facial province with predominant carbonate sedimentation (the Carpathians, Crimea, Ciscaucasia, and southern Middle Asia), and northern facial province with prevalent terrigenous and terrigenous-siliceous sedimentation (the Dnieper-Donets depression, Voronezh antec-

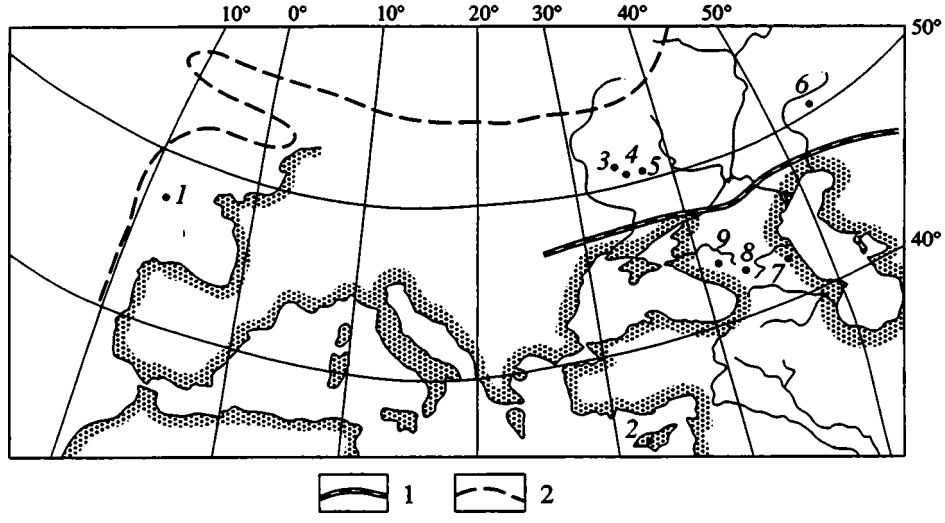

Fig. 1. Location of main sedimentation provinces and studied sections.
(*1*) DSDP Hole 400A; (*2*) Perapedi section; (*3*) Borehole 230; (*4*) Borehole 246; (*5*) Yaruga section; (*6*) Borehole SP-1, (*7*) Rubas-Chai section. 1, boundary between the southern province with carbonate sedimentation and northern province with terrigenous sedimentation; 2, northern boundary of the Tethyan basin in the Eocene.

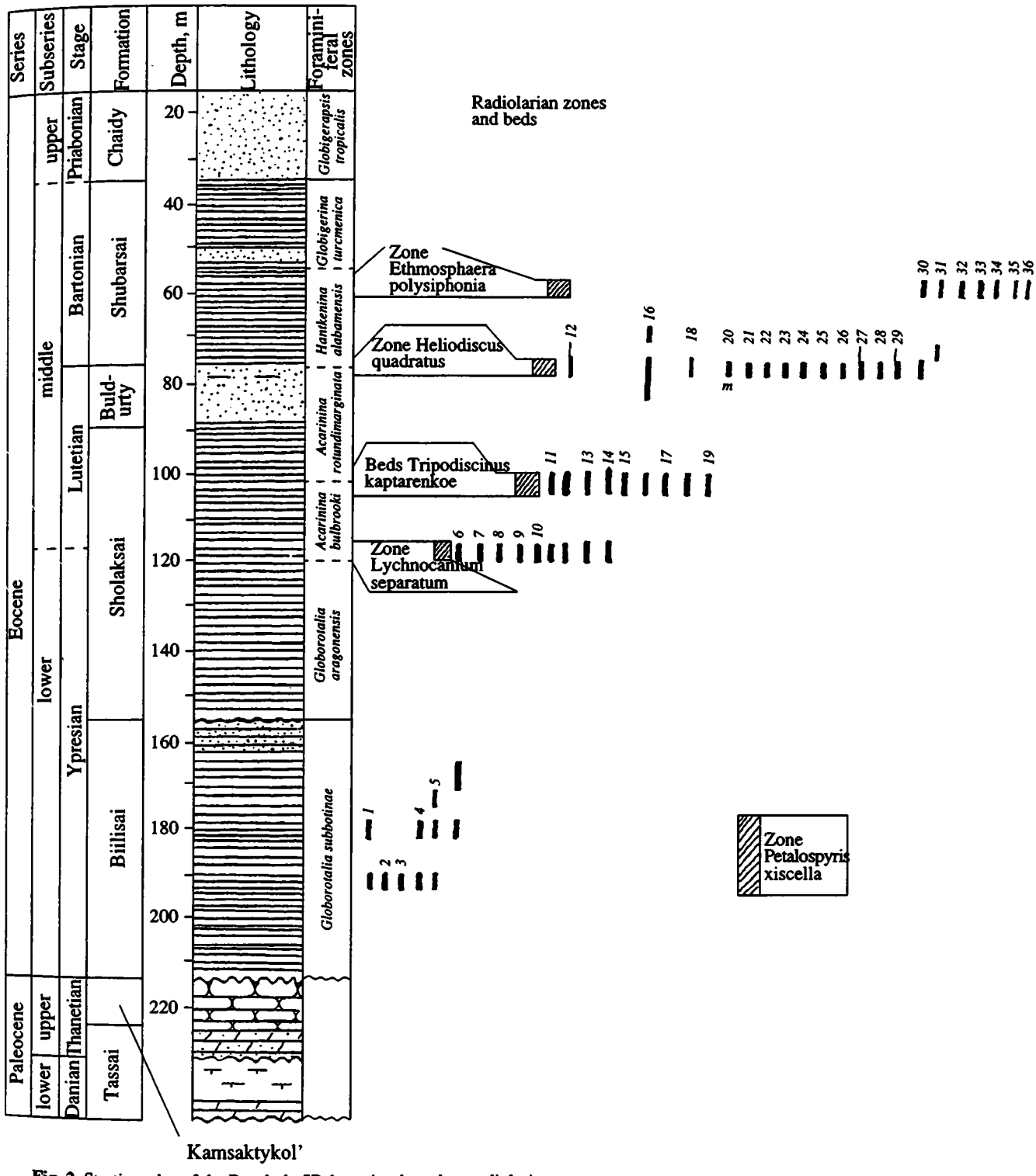

Fig. 2. Stratigraphy of the Borehole SP-1 section based on radiolarians.

(1) Lychnocanoma ponderosa; (2) *Clathrocyclas lipmanae*; (3) *Amphicraspedum prolixum*; (4) *Spongotrochus paciferus*; (5) *Petalospyris fiscella*; (6) *Tripodiscinus* sp.; (7) *Lithostrobus* aff. *turris*; (8) *Stylodyctya torquatum*; (9) *Petalospyris tumidula*; (10) *Heliodiscus heliasterscus*; (11) *Clathrocyclas multiplicatus*; (12) *Lychnocanium separatum*; (13) *Petalospyris compressus*; (14) *Petalospyris triceros*; (15) *Carposphaera magnoporulosa*; (16) *Tripodiscinus tumulosus*; (17) *Tripodiscinus kaptarekoe*; (18) *Tripodiscinus sibiricus*; (19) *Lychnocanium bellum*; (20) *Heliodiscus* cf. *quadratum*; (21) *Podocyrtis papalis*; (22) *Tripodiscinus clavipes*; (23) *Haliomma immensa*; (24) *Clathrocyclas universa*; (25) *Clathrocyclas elegans*; (26) *Theocampe* cf. *reticulum*; (27) *Hwxaconthium perelegans*; (28) *Heliodiscus* aff. *triactis*; (29) *Axoprunum* cf. *chabacovi*; (30) *Theocorys asperum*; (31) *Lophophaena* aff. *sibiricum*; (32) *Artobotrys norvegiensis*; (33) *Tripodiscinus vanus*; (34) *Druppatractus santaennae*; (35) *Porodiscus parvus*; (36) *Etmosphaera polysiphonia*.

lise, Volga region, northern Caspian region, West Siberia, and others) (Fig. 1). Yet, there are still many difficulties regarding the age of the Paleogene radiolarian zones and their correlation with zones defined by other plankton groups, particularly with standard tropical and subtropical scales. This is related to the following circumstances: (1) unfortunately, no sections with a complete and continuous succession of radilolarian zones spanning at least the subseries interval were found on the territory of the former Soviet Union; (2) the Paleogene radiolarian fauna of this region is extremely peculiar in taxonomy, which is explained by its habitat in more or less shallow-water basins; the composition and structure of the assemblages are still insufficiently studied; and (3) in the northern facial province, radiolarians represent the only microfossil group in the Paleogene deposits, or they are accompanied by poor assemblages of other microorganisms, and, therefore, the age determination of host deposits is a difficult problem.

All of these factors predetermine radiolarians from the epicontinental Tethyan areas to be extremely important. The aim of this work was to investigate the taxonomic composition and structure of radiolarian assemblages from seven Paleogene sections in this region, perform their correlation, and to make conclusions on paleoecological settings based on the results. In the southern structural-facial province, the bed-by-bed sampling was performed in the sections of the Kuban' River, Kheu, and Rubas-Chai localities (western, central, and eastern Ciscaucasia, respectively). In the northern structural-facial province, radiolarians were studied in sections of Boreholes 230 and 246 and in the Yaruga locality (Belgorod District, Voronezh anteclise) and Borehole SP-1 (northern part of the northern Caspian region). Data on the taxonomic composition and age of radiolarian assemblages and the lithology of host rocks in the Ciscaucasian sections and Boreholes 230 and 246 were given in earlier publications (Radionova and Khokhlova, 1994 and Radionova *et al.*, 1994, respectively). Therefore, in this paper, I will focus on the lithologic description and stratigraphic distribution of radiolarians only in the Yaruga and Borehole SP-1 sections (Figs. 2 and 3). For comparison, I also used my data on Paleogene radiolarian assemblages from the Biscay Bay (Hole 400A of the Deep Sea Drilling Project) and Cyprus.

STRATIGRAPHIC SUBDIVISION OF BOREHOLE SP-1 (NORTHERN CASPIAN REGION) AND YARUGA (VORONEZH ANTECLISE) SECTIONS BASED ON RADIOLARIANS

The sections are subdivided according to the zonal scale by Kozlova (1990). When the radiolarian assemblage was insufficiently representative for zone recognition, beds with particular radiolarians were defined.

Borehole SP-1

This borehole, located in the northeastern part of the northern Caspian region (Fig. 2), penetrated a continuous succession of Paleogene deposits typical of the temperate-boreal province and consisting of clays and sandstones with an insignificant carbonate admixture. Microfossils from this section were investigated by several researchers (G.E. Kozlova, N.I. Strel'nikova, Z.I. Gleser, and others). Data on the radiolarian and foraminifer distribution were given in the article by Beniamovskii *et al.*, 1990.

The planktonic foraminifer stratigraphy of this section was performed by Ben'yamovskii. The Paleocene clays and marls lack determinable radiolarians. These strata are unconformably overlain by gray–green, somewhat sandy claystones of the Bailisai Formation. In its middle part, the sandy material almost completely disappears, and claystones contain an impoverished but well-preserved radiolarian assemblage associated with planktonic foraminifers of the *Globorotalia subbotinae* Zone (lower Eocene).

Lower Eocene. Ypresian. This interval yielded the assemblage of the *Petalospyris fiscella* Zone, which is characterized by the index species, *Lychnocanoma ponderosa, Amphicraspedum prolixum, Spongotrochus paciferus paciferus*, and *Clathrocyclas lipmanae*. Somewhat higher in the section, *Tripodiscinus sengilensis* and *Helodiscus* aff. *lentis* appear.

The thickness of the Bailisai Formation is about 60 m. Clays of this unit are overlain with slight erosion marks by thin-platted green claystones of the Sholaksai Formation. Its middle and upper parts yielded two radiolarian assemblages, which are correlative with the *Acarinina bulbrooki* and *Acarinina rotundimarginata* foraminiferal zones, respectively.

Middle Eocene. Lutetian. The *Lychnocanium separatum Zone* and *Tripodiscinus kaptarenkoe* Beds were defined in this interval. The assemblage from the former unit is diverse and well preserved. It consists of dominant *Tripodiscinus* sp., *Petalospyris tumidula, P. compressus, P. septenaria*, common *Heliodiscus heliasteriscus, Stylodyctya terquatum*, nasselarians *Lychnocanium separatum, Clathrocyclas multicostatus, C. tamdiensis, Lithostrobus* aff. *turris*, and others. The *Tripodiscinus kaptarenkoe* Beds in the upper part of the Sholaksai Formation contain a diverse and well-preserved assemblage with dominant species of the *Tripodiscinus* genus: *T. tumulosus, T. kaptarenkoe*, and *T. sibiricus*. Also common are small spirids of the *Petalospyris* genus: *P. triceros* and *P. compressus*. Species *Lychnocanium bellum, Carposphaera magnoporulosa*, and *Diplocyclas parcum* are also present.

Higher in the Borehole SP-1 section, the Sholaksai Formation is overlain by sandy deposits of presumably the late Lutetian Buldurtin Formation, which are barren of radiolarians.

The total thickness of the Sholaksai and Buldurtin formations is approximately 80 m. The Buldurtin For-

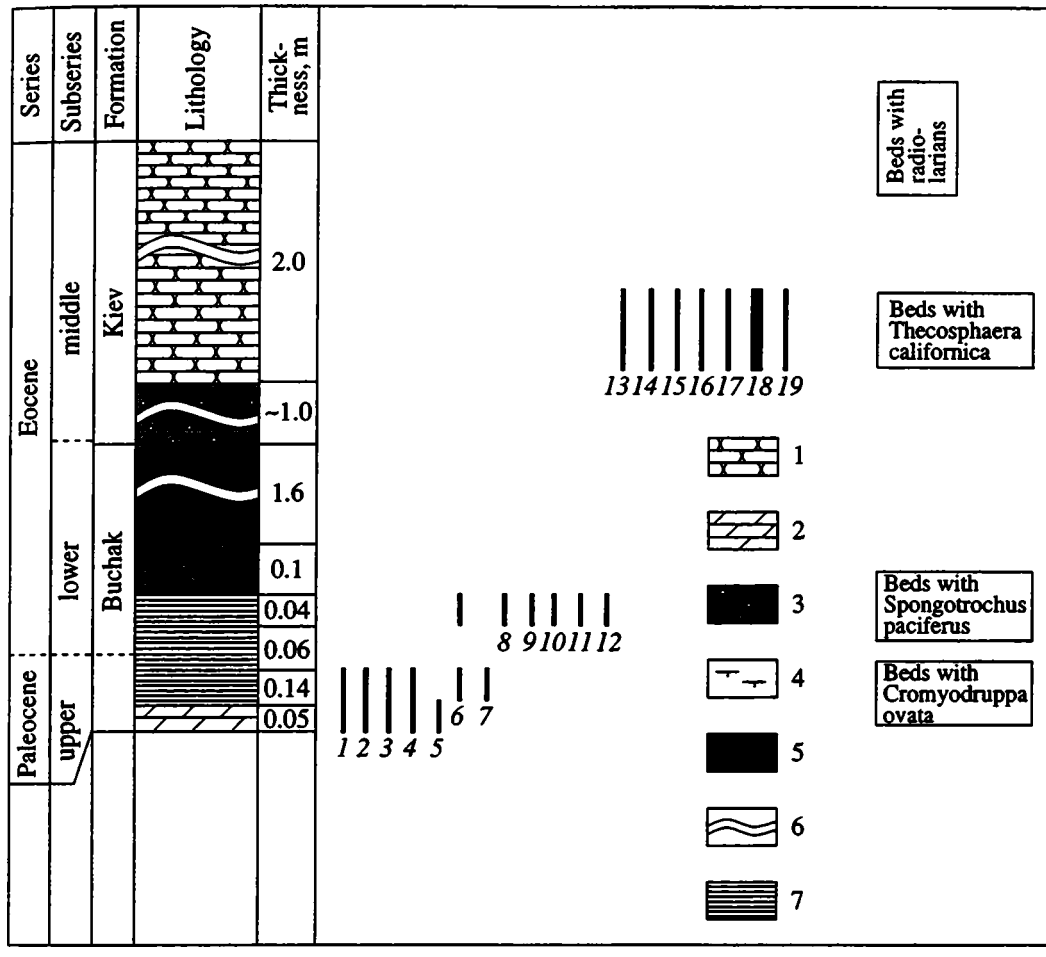

Fig. 3. Stratigraphy of the Yaruga section based on radiolarians.

Lithology: 1, opokas; 2, marl; 3, sandy opokas; 4, chalk; 5, sands; 6, break in column; 7, clays (for Figs. 2 and 3).

Radiolarians: (*1*) *Spongurus*(?) *regularis*; (*2*) *Theocorys* cf. *fiscella*; (*3*) *Cromyodruppa ovata*; (*4*) *Larnocalpis* aff. *smili*; (*5*) *Dictyomitra multicostata*; (*6*) *Stylotrochus harlestonensis*; (*7*) *Cromyiomma riedeli*; (*8*) *Triactis*(?) *angilatum*; (*9*) *Stylotrochus nitidus*; (*10*) *Axoprunum* sp.; (*11*) *Diacanthocapsa* sp.; (*12*) *Spongitrochus paciferus*; (*13*) *Lithomelissa haeckeli*; (*14*) *Thecosphaera californica*; (*15*) *Cyrtophomis* aff. *brevispina*; (*16*) *Botryopera* ex gr. *triloba*; (*17*) *Spongomelissa* aff, *numa*; (*18*) *Plectodiscus circularis*; (*19*) *Stylodyctya targaephormis*.

mation is overlain by the monotonous gray–green clays of the Shubarsai Formation, the lower and middle parts of which contain representative radiolarian assemblages correlative with the *Hantkenina alabamensis* and *Globigerina turkmenica* foraminiferal zones, respectively.

Middle Eocene. Bartonian. Two stratigraphic units were established in the Bartonian deposits: the *Heliodiscus quadratus* and *Ethmosphaera polysiphonia* zones. Diverse and abundant radiolarians from the former zone are mainly represented by the dominant cyrtids *Podocyrtis papalis, Clathrocyclas universa, C. elegans, Theocampe* cf. *reticulum, Theocorys asperum, Lophophaena* aff. *sibirica* and common *Haliomma immensa, Heliodiscus* aff. *triactis, Hexaconthium pereleganS, Axoprunum* cf. *chabakovi*, and *Tripodiscinus clavipes*. Species *Heliodiscus* cf. *quadratus* and *Hexalonche* aff. *diplacantha* are also present. The *Ethnosphaera polysiphonica* Zone is characterized by *Artobotrys norvegiensis, Tripodiscinus vanus, Lophophaena* aff. *sibirica*, abundant *Druppatractus santaennae, Porodiscus parvus, Heterosestrum* sp., *Theocorys asperum*, and rare *Ethmosphaera polysiphonia*. The thickness of the Shubarsai Formation is 40 m.

The Yaruga Section (Southern Flank of the Voronezh Anteclise)

The basal part of the section is represented by a white chalk layer, which is overlain by green, brown, and gray claystones (Fig. 2). The lower and upper parts of the clayey member yielded radiolarian assemblages whose ages can be determined only by comparison with well-known and reliably dated assemblages from

other sections, as the considered section lacks any other reference microfossils.

Paleocene. The *Cromyodruppa ovata* Beds are established in this interval. They contain *Spongurus*(?) *regularis, Cromyodruppa ovata, Theocorys* cf. *phyzella, Larnocalpis* aff. *smili, Dictyomitra multicostata, Cromyomma riedeli*, and *Stylotractus harlestonensis*. The listed species occur in the Marsyat Formation of the Transuralian region (Grigor'eva, 1979), Syzran' Formation of the Volga region (Kozlova, 1984), and Paleocene deposits of the Atlantic (Foreman, 1973) and Ciscaucasia (Borisenko, 1960b). The Paleocene age of this assemblage is believed to be beyond a doubt. The thickness of the Paleocene deposits is 0.18–0.2 m.

The Paleocene clayey deposits are conformably overlain with bluish-gray, locally ferruginous claystones, which yielded an impoverished radiolarian assemblage.

Lower Eocene (?). The *Spongotrochus paciferus* Beds contain a radiolarian assemblage of *Triactis* aff. *angustum, Stylotrochus harlestonensis, S. nitidus, Axoprunum* sp., *Diacanthocapsa* sp., and *Spongotrochus paciferus*. As these species are all of a wide stratigraphic range, the assemblage can only be arbitrarily dated as the early Eocene. This age estimate is supported by the cooccurrence of species *Spongotrochus paciferus, Stylotrochus harlestonensis*, and *S. nitidus*, which are widespread in the lower Eocene sediments of the World Ocean.

The thickness of presumably lower Eocene claystones is 0.05–0.1 m. They are overlain without visible unconformity by thin-platted sandy opokas, which, in turn, are replaced upsection by an opoka-like sandstone member bearing glauconite in its basal part. This 2.7-m-thick member lacks microfossils. Higher in the section, light-gray opokas occur; the lower part of this opoka member contains radiolarians.

Middle Eocene. Based on radiolarians, the *Thecosphaera californica* Beds were defined in the opoka member. Their radiolarians are more diverse and better preserved as compared with two previous assemblages. The assemblage from this interval is dominated by *Plectodiscus circularis*. Also common are *Lithomelissa haeckeli, Thecosphaera californica, Cyrtophormis* aff. *brevispina, Botryopera* ex gr. *triloba, Stylodyctya targaephormis*, and *Spongomelissa* aff. *numa*. Although this assemblage is devoid any zonal forms, the listed species allow it to be confidently dated as the middle Eocene because they were recorded in coeval deposits of the Russian platform, Atlantic, and California (Campbell and Clark, 1945; Petrushevskaya and Kozlova, 1972; Radionova *et al.*, 1994).

The visible thickness of the opoka member is 2 m. In its upper part, the unit is sandy and barren of radiolarians.

CORRELATION OF STUDIED SECTIONS BASED ON RADIOLARIANS

The age estimates of radiolarian assemblages from the seven analyzed sections were correlated, when it was possible, with the data on other planktonic groups (nannoplankton, foraminifers, diatoms, and/or silicoflagellates) occurring in the same samples (Fig. 4). They were studied by N.G. Muzylev, V.N. Ben'yamovskii, T.V. Oreshkina, and E.P. Radionova, respectively. In all of the studied sections, radiolarians occur only at certain levels, which correspond, as a rule, to well-known eustatic transgressions. In the northern province of the Tethyan epicontinental basins, radiolarian assemblages are confined to the base of sedimentary cycles, whereas in the southern province, their position is not so clear (Figs. 2 to 4).

In the sections under consideration, deposits with radiolarian assemblages are confined to the following time intervals: (1) the late Thanetian (Kheu, Kuban', and Yaruga sections); (2) early Ypresian (Kheu, Rubas-Chai, Kuban', and Borehole SP-1 sections); (3) Ypresian (Kheu, Kuban', and Borehole SP-1 sections); (4) middle Lutetian (Kuban', Kheu, Rubas-Chai, and Borehole SP-1 sections); (5) late Lutetian (Boreholes 230 and SP-1, Rubas-Chai, and Yaruga sections); and (6) early Bartonian. In the southern province, with its predominantly carbonate sedimentation, the undivided Bartonian interval is distinguished. During this interval, clays of the Kuma Formation with common radiolarians (Moksyakova, 1972; Lipman, 1953; and others) accumulated throughout the southern part of the former Soviet Union. In the northern province, there are two intervals characterized by radiolarians: (6a) the early Bartonian and (6b) late Bartonian (sections of the Voronezh anteclise and Boreholes 230, 246, and SP-1 in the northern Caspian region). Late Eocene (Priabonian) radiolarians were not found in the studied sections. In the southern part of this region, one more level with radiolarians can be defined: (7) the early Rupelian (Kuban', Kheu, and Rubas-Chai sections). In the northern province, this level is not traced. The characteristics of radiolarian assemblages from the above-mentioned time intervals follow below.

1. The late Thanetian radiolarian fauna is impoverished and poorly preserved. Most of the species recorded in the Ciscaucasia and Voronezh anteclise (Yaruga section) were previously reported from the Syzran' Formation of the Volga region, Marsyat Formation of the Cis-Urals region, and Talitsa Formation of West Siberia, as well as from the Atlantic and Creek Formation of Missouri (Kozlova, 1984; Grigor'eva, 1979; Foreman, 1973). These are *Cromyodruppa ovata, Spongurus*(?) *regularis, Larnocalpis smili, Stylotrochus harlestonensis, Thecosphaera rotunda, Theocorys phyzella, Lamptonium pennatum*, and others. The assemblage mostly consists of the species *Ommatogramma, Spongodiscus, Cromyodruppa, Petalospyris, Cinosphaera*, and sometimes *Lychnocanoma*

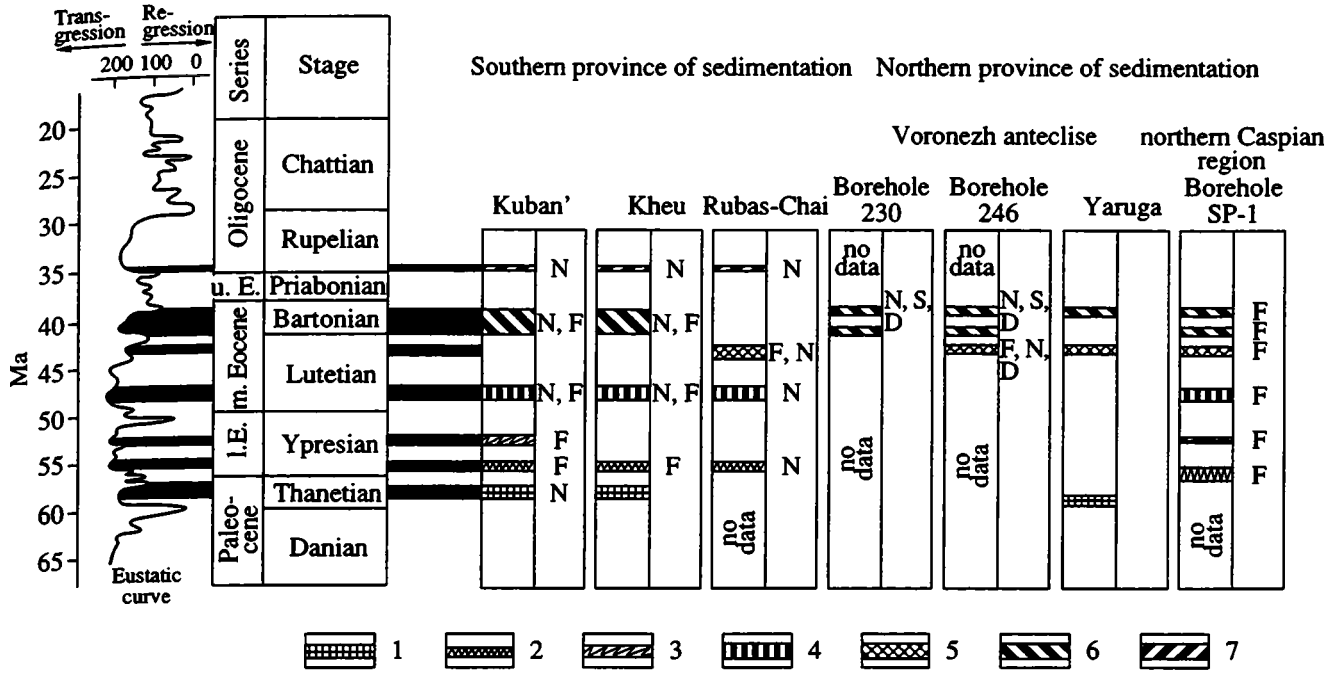

Fig. 4. Correlation of stratigraphic intervals containing radiolarins with the eustatic curve.
Intervals: 1, Thanetian; 2, lower Ypresian; 3, upper Ypresian; 4, middle Lutetian; 5, upper Lutetian; 6, Bartonian; 7, lower Oligocene.
N, nannoplankton; F, foraminifers; D, diatoms.

and *Buryella* genera. Stratigraphically important tropical *Becomiforma* and *Theocotyle* genera are absent from this assemblage.

2 and 3. Both early Eocene radiolarian assemblages in the considered areas are of low diversity, but commonly well preserved. Data by Krasheninnikov (1960) and Borisenko (1960a) on the lower Eocene of western Ciscaucasia and by Kozlova (1984) on the coeval deposits of the Volga midstream region, along with my data on the eastern Ciscaucasia and northern Caspian region (Borehole SP-1), shows the presence in these assemblages of many species in common with those from the lower Eocene sediments of oceanic basins, particularly the tropical and subtropical Atlantic. These are *Buryella tetradica, B. clinata, Phormocyrts hirsuta hirsuta, Lychnocanoma ponderosa, Amphicraspedum prolixum, Petalospyris fiscella, Sponotrochus paciferus paciferus*, and others. Evidently, the lower Eocene radiolarian fauna from the peripheral Tethyan basins in the former Soviet Union territory can be directly correlated with the standard radiolarian zones (Riedel and Sanfilippo, 1978).

4, 5, and 6. Quite a different situation exists with both the Lutetian and Bartonian radiolarian faunas. Whereas the Lutetian assemblages from the southern province (eastern Ciscaucasia) include many species also characteristic of the oceanic tropical belt (for instance, *Dictyopora mongolfieri, Calocycloma ampulla, Phormocyrtis reticula, Lychnocanium bellum, Sethocyrtis babylonis, Stylatractus pachystylus, Podocyrtis papalis*, and others; Radionova and Khokhlova, 1994), and the *Dictyopora mondolfieri* Zone of the standard radiolarian scale can be defined there, the radiolarian fauna composition in the northern province is essentially different. In the Lutetian and Bartonian assemblages from the Voronezh anteclise and northern Caspian regions, dominant forms are temperate-boreal species, which commonly occur in coeval deposits of West Siberia (Kozlova and Gorbovets, 1978), the Norwegian–Greenland basin (Kozlova and Petrushevskaya, 1979), and Dnieper-Donets (Tochilina, 1975) and Voronezh (Radionova *et al.*, 1994) regions. However, the Bartonian radiolarian assemblages from the northern Caspian region contain rare tropical and subtropical elements, whereas the coeval associations from the Voronezh anteclise show only the presence of *Axoprunum chabakovi* and *A. visendum*, which immigrated from the southern province, where the diverse well-known radiolarian fauna is characteristic of the Kuma Formation brown clays widespread throughout the Ciscaucasia and Middle Asia regions.

PALEOECOLOGY OF THE LUTETIAN BASINS IN EASTERN CISCAUCASIA AND NORTHERN CASPIAN REGIONS BASED ON RADIOLARIANS

The middle Eocene radiolarians are most widespread and representative in the considered regions. Among other investigated assemblages, they are more reliably dated and traced over a large territory. Accord-

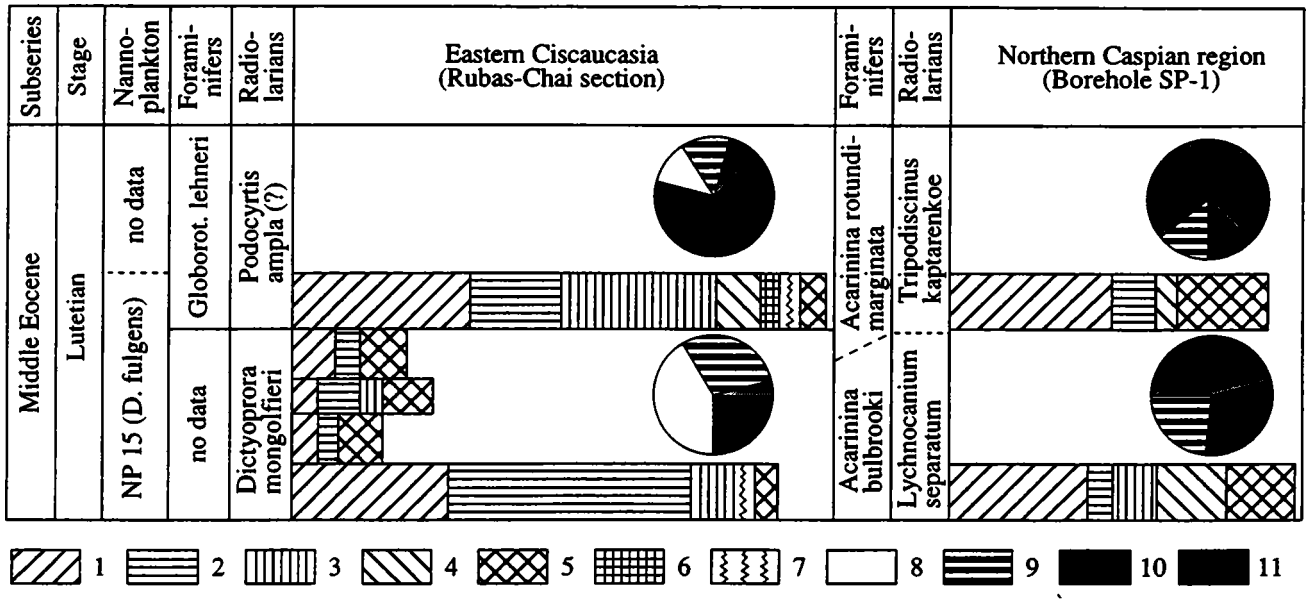

Fig. 5. Structure of middle Eocene radiolarian assemblages in the eastern Ciscaucasia (Rubas-Chai section) and northern Caspian region (Borehole SP-1).

1, Cyrtida; 2, Actinommiidae; 3, Spongodiscidae; 4, Phacodiscidae; 5, Spyrida; 6, Liteliidae; 7, Coccodiscidae; 8, tropical-subtropical; 9, subtropical; 10, cosmopolitan; 11, temperate-boreal and boreal.

ingly, radiolarians from the southern areas of Russia and adjacent regions serve as the basis for paleoecological interpretations. The relative proportions of their specimens in higher taxa (families and, sometimes, suborders) were estimated for two radiolarian assemblages from the middle and upper Lutetian intervals of the Rubas-Chai (eastern Ciscaucasia) and Borehole Sp-1 (northern Caspian region) sections.

When interpreting the data, I used the method by Kruglikova (1979, 1981) along with the data available on radiolarian ecology (Anderson, 1981; Boltovskoy, 1987; and others). The Lutetian deposits of the abovementioned sections are also very important for such interpretations, because the lithology of these sections can be considered as monotonous throughout the Lutetian interval; this eliminates, to some extent, the effect of selective preservation of radiolarians in different facies, which distorts the assemblage initial composition.

The Rubas-Chai Section

The Middle Lutetian (the NP 15 or *Nannotetrina fulgens* Zone). In this interval, radiolarians are abundant, diverse, and well preserved. Actinomiids are dominant; cyrtids also play an important role. The low relative content of spongodiscids and small spyrids is a peculiar feature of the assemblage (Fig. 5). It mostly consists of tropical and subtropical species; approximately a quarter of the assemblage is represented by cosmopolitan forms. Actinomiids show well-developed spines. It is known (Matsuoka and Andersen, 1992) that actinomiids lose their spines because of silica dissolution when the habitat environments become abnormal. The taxonomic composition and structure of the considered assemblage are very similar to those in coeval assemblages from the Biscay Bay (Hole 400A), which points to their development in a warm oceanic basin with normal salinity during the middle Lutetian transgression.

Upsection, the structure of the radiolarian assemblage abruptly changes. The lower diversity, poorer preservation, predominance of small spirids, and reduced role of cyrtids and actinomiids are the principal features of this fauna. The similar structure of the assemblage is generally typical of the regressive phase.

The upper Lutetian (the foraminiferal *Morozovella lehneri* Zone). The assemblage is very diverse. Cyrtids and spongodiscids are predominant, whereas the share of actinomiids is insignificant. Small spirids and coccodiscids are subordinate constituents of the assemblage as well. Also present are litheliids, which are generally considered to represent the boreal assemblages. As compared to older associations, the share of cosmopolitan forms increases, whereas that of tropical and subtropical species diminishes. Boreal elements occur sporadically. The high intraspecies variability of spongodiscids and coccodiscids is a peculiar feature of the assemblage. According to Kruglikova's (1990) data, this generally testifies to the fauna habitat in epicontinental marine environments. There are also evidences that some spongodiscids show a certain tolerance to salinity changes; according to Matsuoka and Anderson (1992), they can survive salinity as low as 27‰. In addition, cosmopolitan zooplankton species are generally

more euryhaline than species with a narrower geographic range. The high taxonomic diversity of the assemblage without the distinct prevalence of certain (one or two) species suggests that it dwelt in open marine environments. Considering these facts, one can argue that the late Lutetian radiolarian fauna of eastern Ciscaucasia developed in a basin with reduced salinity, which was located at the oceanic shelf. Geographically, the basin was probably confined to the northern part of the subtropical belt.

The Borehole SP-1 Section

The middle Lutetian (the foraminiferal *Acarinina bulbrooki* Zone). Radiolarians are well preserved and diverse, with cyrtids being the dominant group. Also abundant are phacodiscids and small spirids. The latter group shows high intraspecies variability. Spongodiscids and, particularly, actinomiids play an insignificant role. In general, the assemblage predominantly consists of boreal and temperate-boreal elements, with a complete absence of tropical species. The number of cosmopolitan forms is also high. The assemblage structure suggests its habitat basin to be of the shallow-water epicontinental type located in the shelf area of the southern part of the boreal province. Upsection, deposits corresponding to the regressive stage yielded only small spirids and rare spongodiscids.

The late Lutetian (the foraminiferal *Acarinina rotundimarginata* Zone). The radiolarian assemblage is more diverse than in the middle Lutetian interval. Predominant are cyrtids and small spirids. The number of actinomiid species remains almost equal to that at the beginning of the middle Lutetian, while the share of spongodiscids and coccodiscids diminishes. The predominance of the cyrtid forms of the Clathromitridae family (particularly, those of the *Tripodiscinus* genus, which is widespread in the middle Eocene sequences of the boreal realm) is characteristic of the considered assemblage. Starting from this stratigraphic level, the distinct similarity between the radiolarian associations of southern Russia, the Volga region, and Norwegian basin became evident. It is quite obvious that in contrast to the middle Lutetian, the late Lutetian basin was shallower, more isolated from the main oceanic basin, and located in the boreal province. The influence of early to late Lutetian cooling is reflected in the taxonomical composition of radiolarian faunas in both eastern Ciscaucasia and the northern Caspian region.

CONCLUSION

In Paleogene sections of southern Russia and adjacent regions, radiolarians occur at certain stratigraphic levels that correspond to well-known eustatic transgressions. The taxonomic composition of coeval assemblages is different in separate areas of the considered region, thus reflecting the paleoclimatic and paleoceanographic peculiarities. In the early Paleocene, the relatively low and laterally uniform temperature of surficial water masses was responsible for the taxonomic uniformity of radiolarian fauna throughout the North Tethyan region. Progressive warming during the Paleocene culminated in the Cenozoic optimum at the beginning of the early Eocene. Along with the water exchange between the Tethys, Indian, and Atlantic oceans and the absence of a distinct latitudinal temperature gradient, this promoted the development of the taxonomically uniform radiolarian fauna in epicontinental marginal basins of southern Russia and the tropical part of the Tethys during the early Eocene. The cooling at the beginning of the middle Eocene resulted in the latitudinal differentiation of water masses, which affected the radiolarian fauna composition. The formation of two paleogeographic provinces—tropical and temperate-boreal—with their circulation systems, temperature regimes, and sedimentation patterns led to the gradual development of taxonomically and structurally different radiolarian faunas.

ACKNOWLEDGMENTS

The work was financially supported by the International Program *Peritethys* (project 95-16).

Reviewer I.A. Basov

REFERENCES

Anderson, O.R., Radiolarian Fine Structure and Silica Deposition, *Silicon and Siliceous Structures in Biological Systems*, Simpson, T.L. and Volcani, B.E., Eds., New York: Springer, 1981, pp. 347–349.

Boltovskoy, D., Sedimentary Record of Radiolarian Biogeography in the Equatorial to Antarctic Western Pacific Ocean, *Micropaleontology*, 1987, vol. 33, no. 3, pp. 267–281.

Beniamovskii, V.N., Segedin, R.A., Akopov, T.R., *et al.*, New Paleocene and Eocene Formations in the North Caspian Region, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1990, vol. 65, issue 5, pp. 68–76.

Borisenko, N.N., Radiolarians from the Lower and Middle Eocene of the Western Kuban' Region, *Tr. Vses. Nauchno-Issled. Inst. Nefti Gaza*, 1960a, issue 4, pp. 219–232.

Borisenko, N.N., New Paleocene Radiolarians from the Kuban' Region, *Tr. Vses. Nauchno-Issled. Inst. Nefti Gaza*, 1960b, issue 4, pp. 199–208.

Campbell, A. and Clark, B.L., Radiolaria from the Kreyenhagen Formation Near Los Banos, California, *Mem. Geol. Soc. Amer.*, Baltimore: Weverly, 1945, no. 10.

Foreman, N.P., Radiolaria of Leg 10 with the Systematics and Ranges for the Families Amphinydacidae, Artostroboliidae, and Theoperidae, in *Initial Rep. Deep Sea Drill. Project*, Washington: US Gov. Print. Office, 1973, vol. 10, pp. 407–469.

Gorbunov, V.S., *Radiolarii srednego i verkhnego eotsena Dneprovsko-Donetskoi vpadiny* (Middle and Upper Eocene Radiolarians from the Dnieper-Donets Depression), Kiev: Naukova Dumka, 1979.

Grigor'eva, A.I., Radiolarians from Upper Cretaceous and Paleogene Deposits of the Urals Eastern Slope and Cis-

Urals, *Tr. Inst. Geol. Geofiz., Ural. Nauchn. Tsentr, Akad. Nauk SSSR*, 1975, issue 119, pp. 102–109.

Kozlova, G.E., Zonal Stratigraphy of the Boreal Paleogene Based on Radiolarians, in *Morfologiya, ekologiya i evolutsiya radiolarii* (Morphology, Ecology, and Evolution of Radiolarians), Leningrad: Nauka, 1984, pp. 196–210.

Kozlova, G.E., Philogenetic Research as the Basis for Elaboration of the Zonal Scale of the Boreal Paleogene Based on Radiolarians, in *Radiolarii v biostratigrafii* (Radiolarians in Biostratigraphy), Sverdlovsk: Ural. Nauchn. Tsentr, Akad. Nauk SSSR, 1990, pp. 70–81.

Kozlova, G.E. and Gorbovets, A.I., Radiolarians from Upper Cretaceous and Upper Eocene Deposits of of the West Siberian Lowland, *Tr. Vses. Nauchn.-Issled. Geol.-Razved. Inst.*, 1966, issue 248.

Kozlova, G.E. and Petrushevskaya, M.G., Stages in Radiolarian Evolution in the Norwegian–Greenland Basin, in *Istoriya mikroplanktona Norvezhskogo morya* (Microplankton Evolution in the Norwegian Sea), Leningrad: Nauka, 1979, pp. 71–85.

Krasheninnikov, V.A., Some Radiolarians from the Lower and Middle Eocene of the Western Ciscaucasia, *Tr. Vses. Nauchno-Issled. Inst. Nefti Gaza*, 1960, issue 16, no. 3, pp. 271–301.

Kruglikova, S.B., Characteristic Features in the Distribution of Radiolarians (Higher Taxa) in Sediments of the Pacific and Marginal Seas, in *Iskopaemye i sovremennye radiolarii* (Fossil and Recent Radiolarians), Leningrad: Nauka, 1979, pp. 42–52.

Kruglikova, S.B., Some Ecological and Distributional Features of Recent and Cenozoic Radiolarians, in *Sistematika, evolyutsiya i stratigraficheskoe znachenie radiolarii* (Systematics, Evolution, and Stratigraphic Significance of Radiolarians), Moscow: Nauka, 1981, pp. 118–139.

Kruglikova, S.B., Radiolarians as Indicators of Some Paleoenvironmental Parameters, in *Radiolarii v biostratigrafii* (Radiolarians in Biostratigraphy), Sverdlovsk: Ural. Nauchn. Tsentr, Akad. Nauk SSSR, 1990, pp. 92–106.

Lipman, R.Kh., Eocene Radiolarians of the Kyzyl-Kum Region, *Tr. Vses. Geol. Inst.*, 1950, issue 1, pp. 51–66.

Lipman, R.Kh., Materials on Studies of Radiolarians from Western Areas of Middle Asia, *Ezheg. Vses. Paleontol. O-va*, 1953, vol. 14, pp. 135–162.

Matsuoka, A. and Anderson, O.R., Experimental and Observational Studies of Radiolarian Physiological Ecology. 5: Temperature and Salinity Tolerance of *Dictyocoryne truncatum*, *Mar. Micropaleontol.*, 1992, no. 19, pp. 299–313.

Moksyakova, A.M., *Bodrakskii yarus Turanskoi plity (stratigrafiya, radiolyarii i paleozoogeografiya)* [Bodrakian Stage of the Turan Platform (Stratigraphy, Radiolarians and Paleozoogeography)], Moscow: Nedra, 1972.

Petrushevskaya, M.G. and Kozlova, G.E., Radiolaria: Leg 14, Deep Sea Drilling Project, in *Initial Rep. Deep Sea Drill. Project*, Washington: US Gov. Print. Office, 1972, vol. 14, pp. 495–647.

Radionova, E.P. and T.V., Khokhlova, Episodes of Biogenic Siliceous Sedimentation in the Northern Caucasus and Adjacent Regions of the Tethyan Province, *Stratigr. Geol. Korrelyatsia*, 1994, vol. 2, no. 5, pp. 161–169.

Radionova, E.P., Oreshkina, T.V., Khokhlova, I.E., and Ben'yamovskii, V.N., Eocene Deposits of the Northeastern Flank of the Dnieper-Donets Depression (Zonal Stratigraphy and Cycle Analysis), *Stratigr. Geol. Korrelyatsia*, 1994, vol. 2, no. 6, pp. 85–102.

Resheniya MSK. Byulleten' regional'noi mezhvedomstvennoi komissii po tsentru i yugu Russkoi platformy, Vyp. 1 (Resolutions of ISC. Bulletin of Regional Interdepartmental Commisision on the Central and Southern Russian Platform), Moscow: Rosgeolfond, 1989, issue 1.

Riedel, W.R. and Sanfilippo, A., Stratigraphy and Evolution of Tropical Cenozoic Radiolarians, *Micropaleontology*, 1978, vol. 24, no. 1, pp. 61–96.

Tochilina, S,V., Biostratigraphic Characteristics of Paleocene–Eocene Deposits of the Voronezh Anteclise. Systematics and Stratigraphic Significance of Radiolarians, *Tr. Vses. Geol. Inst., Novaya Ser.*, 1975, vol. 226, pp. 78–84.

Zagorodnyuk, V.M., Radiolarians from Eocene Deposits of the Lower Don River Region, *Cand. Sci. (Geol.–Mineral.) Dissertation*, Novocherkassk: Polytechnical Inst., 1970.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 371–377. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 62–68.
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The Eocene–Early Miocene Evolution of Plant Assemblages and Large Mammals in the Extratropical Asia

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Received October 15, 1994

Abstract—The problems considered concern the Eocene–early Miocene evolution of vegetation cover and large mammals in extratropical Asia. As a rule, the evolutionary stages revealed coincided with certain moments of ecological rearrangement related to geologic and paleogeographic events. The narrow-leaved xeromorphic flora of the Bartonian marked the closure time of an open marine strait, which connected the Tethys and Arctic basins. This was a time of intense radiation of phytophagous mammals and coeval development of endemism of Asiatic ungulates. The mesoxerophilic flora of evergreen and deciduous plants, some of which later became basic elements of the Turgai flora, existed at the prosperity time of brontotherioid and tapiroid faunas. Giant Indricotheriinae were widespread at the mid-Oligocene time, when the climate in expanded open landscapes became more arid. Formation of deciduous flora of the Turgai ecotype was completed under conditions of the humid, moderately warm climate of the late Oligocene. Remains of large mammals of this time are very scarce.

Key words: *Eocene, Oligocene, Miocene, Brontotheriidae, Indricotheriinae, Gomphotherium, Turgai flora, forest and marshy vegetation, shrubs, herbs, climate*

INTRODUCTION

One of the basic changes in the evolution of mammals was their adaptation to vegetarian food and the appearance of respective morphological features. Early stages of transition in the diet from the sarcophagous (insects, worms, mollusks, some other invertebrates, and probably, small vertebrates) to euryphagous, and then to the phytophagous way of life were associated with changes in the behavior stereotype, and affected the morphological structures of the entire skeleton.

In the middle latitudes of extratropical Asia, the evolution history of mammalian fauna from the second half of the middle Eocene to the middle Miocene inclusive is divisible into three intervals coeval to the most prominent geological and paleogeographic events. The paleogeographic reorganization included changes in the shoreline configuration and spatial position of an epicontinental basin, which occupied the southern provinces of the Russian platform and extended to the eastern flank of the Turgai depression. Going northward along this depression, the basin stretched into the West Siberian lowland. In the southeast, it was approaching Central Asia. The rearrangement of the drainage system and activation of erosion at the moments of lowered base level and marine regression influenced the character of vegetation and landscapes, and transformed biotopes.

STAGES IN EVOLUTION OF PLANT ASSEMBLAGES AND LARGE MAMMAL COMMUNITIES

Middle Eocene

In the second half of the middle Eocene, the epicontinental marine basin—a branch of the Tethys—was most likely disconnected from the Polar seas. During the Bartonian time, it was closed near the shoreline, or somewhat further to the north, inside the present Arctic Ocean. Owing to this, the middle and even high latitudes of extratropical Asia were for a long time under the influence of warm Tethyan currents, which were responsible for the subtropical climate of western Asia, with hot and dry summers and distinctly seasonal rainfall during winter periods, as it is recorded in plant remains. The evergreen sclerophyllous oak–laurofoliate and subordinate mixed forests with some admixture of mesophilous deciduous plants represented the main type of zonal vegetation in elevated plains around the coastal lowland during the latest middle and earliest late Eocene. Conifers, the *Pinus* forms in elevated areas and Taxodiaceae in more humid coastal zones, were notable elements of forests only in northern West Siberia. Palynoflora from the lower beds of the Chegan Formation and the Takyrsor leaf flora (Makulbekov, 1972) suggest that forest vegetation in central and southern Kazakhstan and southern West Siberia was dominated by *Sequoia*, Cupressaceae, narrow-leaved xeromorphic *Quercus tacyrsorianus, Ilex*, Ericaceae (*Leocothoe*),

laurofoliate species, diverse Leguminosae, Proteaceae, and Myrtaceae. Associated forms were represented by Juglandaceae, Moraceae, Araliaceae, and Anacardiaceae. The meandering river valleys open toward the coastal plain were overgrown with narrow-leaved *Populus*, sclerophyllous *Liquidambar, Myrica, Comptonia*, and other microfoliate shrubs (Ericaceae, Myrtaceae, *Leptespermites*). The assemblage of aquatic and riparian plants consisted of *Nelumbo*, Nymphae, Zingiberaceae, Cyperaceae, *Arundo*, and aquatic ferns (*Agolla, Salvinia*). Halophytes (Chenopodiaceae, *Nitraria, Ephedra*, and some Ericaceae) grew on saline soils of coastal plains periodically flooded by sea. These communities also included some Palmae (*Sabal* and *Trachycarpus*). Taxodiaceae and *Nyssa* were not as abundant as at the commencement of Oligocene. Thin savanna-like forests were widespread on elevated plains.

River valleys and coastal lake areas in the Zaisan depression remote from the sea were habitats of Juglandaceae, Myricaceae, *Dryophyllum, Liquidambar, Ailanthus, Alnus*, and narrow-leaved poplars, the ancestors of current *Populus* ex. sect. *Turanga*.

The Eocene was a time of fast radiation of phytophagous mammals displaying the growing endemism degree of Asiatic ungulate assemblages. The latter became more diverse, and some of their representatives (Brontotheriidae and Indricotheriinae) were quite huge. Originally, many phytophagous mammals had low-crowned teeth in the cheek row, and this indicates, as for current ungulates, that the animals fed on the soft parts of shrubs and trees (leaves and shoots). Animals with hypsodont teeth later adapted to eat more crude parts of plants and some types of fauna. The Eocene perissodactyls demonstrate a fine example of adaptation to vegetarianism. They experienced changes in the dental system, facial part of the cranium, and limbs. The most evident changes are recorded in the incisor-canine dental complex (the important indicator of taking in and biting off the vegetation); in the premolar dental row, which serves to preliminary cut the seized plants; and in the molar row responsible for rumination. Diverse forms of Eocene ungulates from Central Asia intensively utilized the vegetable food. Marshy coastal areas here were inhabited by brachypodous brontotheres, who had canine teeth and incisors of conic form, and low-crowned teeth in the cheek row. Most of the hyracodonts likely fed on leaves and shoots; they had a light skeleton adapted to their running behavior, and a dental system with narrow-crowned incisors, conic canine teeth, and a long row of molars. The diversity of Eocene tapirids from Asia (five families and more than 12 genera) is indirect evidence of the high productivity of vegetal communities of that time (see table).

Some of tapirid families (Helaletidae, Lophealetidae, and Tapiridae) display changes in the cranium structure related to the reorganization of the dental system and followed by the development of the proboscis, or mobile upper lip, which resulted in the contraction of nasal bones.

The dentition specialization of tapirids developed in two directions: first, it resulted in the appearance of bilophodont teeth (families Helaletidae, Deperetellidae, and Tapiridae); and second, in the formation of lophodont teeth at the expense of a longer outer wall of the ectoloph (Lophialetidae and Lophodontidae). These features make the tapirids similar to other Ceratomorpha. The larger mastication surface in the first group was attained by means of molarization of premolar teeth, and therewith, the mostly transverse ridges were responsible for the friction between upper and lower teeth. In the second group with lophodont-type dentition, the mastication surface was sometimes enlarged at the expense of reduction of the foremolar row (especially typical of Breviodontinae) and greater length of the molar row. In this case, upper and lower teeth are rubbed by mastication also along the ectoloph (this is very characteristic of Rhodopaginae).

Some Eocene families of Asiatic tapirids adapted to running and residing in open landscapes. Representatives of Lophialetidae, Deperetellidae, Helaletidae, and Colodontinae also demonstrate a remarkable growth of distal limb divisions and reduction of their lateral dactyls (down to three).

Late Eocene

During the second half of the late Eocene, the climate in all of the regions was cooler, with alternating arid and humid phases. Fagaceae and Lauraceae became less frequent in the zonal forest spectrum, being replaced by gymnosperms (*Pinus, Metasequoia*, and Cupressaceaee) on the one hand, and by broad-leaved mesophylls (*Diospyros, Morus, Liquidambar, Corylopsis, Fothergilla, Hamamelis, Nyssa, Tilia*, and *Ailanthus*) on the other. These forests were rich in lianas represented principally by *Vitis* and *Ampelopsis* genera. The taxonomic spectrum of grass and shrub forms was also wider, and included, according to palynological data, stunted Myrtaceae, Rosaceae, Urticaceae, Liliaceae, and Umbelliferae, in addition to Lauraceae and Myricaceae. *Nelumbo, Potomogeton, Arundo*, and *Zingiberopsis* became even more abundant in aquatic and riparian communities.

Early Oligocene

Because of cooling, which was especially pronounced at the beginning of the Oligocene time, flora was progressing toward the mesophilous type. The climate of that time was still seasonal, especially in the central and southern areas of Kazakhstan. The longer (dozens and few hundreds ka) and more humid phases alternated with periods of relative aridization. The irregular rhythm of climatic fluctuations was very favorable for transformations of vegetative cover, and accelerated the flora evolution. Against this back-

Types of mammalian fauna and composition of vegetal communities of the middle Eocene–early Miocene period in Kazakhstan and adjacent regions of extratropical Asia

Age, Ma	Series	Subseries	Stage	Types of mammalian fauna	Climate		Transgressive ⊕ and regressive ⊖ phases	Marshy brushwood and riparian forests	Forests in flat lowlands	Vegetation of elevated plains, thin and savanna-like forests, brushwood (open and semiclosed vegetation)
20	Miocene	Lower	Burdigalian	*Gomphotherium*	Semiarid	Moderately warm, approaching subtropical	⊕	Rhinocerotids (*Brachypotherium*), poboscideans (*Platybelodon*), and suids (*Anthracotherium*). Riparian *Populus–Fraxinus–Ulmus* forests with Rosaceae in underwood	Proboscideans (*Serridentinus*) and ruminants (*Stephanocemas, Dicrocerus, Lagomeryx*). Foliar forests (Juglandaceae, Ulmaceae, *Castanea*) with *Ilex*, Rosaceae, and Leguminosae in underwood	Mastodonts (*Gomphotherium*), perrisodactyls (*Anchitherium*), rhinocerotids (*Aceratherium*), and ruminants (Bovidae). Microfoliate *Quercus–Zelkova* forests with Leguminosae. *Paliurus–Ziziphus* shibliak with Leguminosae, Myrtaceae, and Rosaceae. Bush–herbaceous coenoses in restricted areas
21.5			Aquitanian		Semihumid					
				Paraceratherium		Moderately warm	⊖ ⊕	Rhinocerotids (*Aprotodon*), ruminants (*Lophiomeryx*), and suids (*Anthracotherium*). Riparian forests of narrow-leaved *Populus* ex sect. *Turanga*, with admixture of *Salix, Alnus*, and Leguminosae. Cyperaceae, *Typha, Sparganium*, and *Arundo* along the river banks	*Chalicotherium borissiaki*, rhinocerotids (*Protaceratherium*), and ruminants (*Eumeryx*). Impoverished mixed forests of the Turgai type with *Fagus, Carpinus*, Frangulaceae, *Acer*, and *Anacardium–Alnus* brushwood	Indricotheriinae (*Paraceratherium*), hyracodonts (*Tenizia*), and ruminants (*Prodremotherium*). Pine and oak forests with Rosaceae and Leguminosae in underwood. Thin savanna-like forests and shibliak consisting of Anacardiaceae, Leguminosae, and Ericaceae
25		Upper	Chattian		Humid			Anthracotheres. *Taxodium–Populus* and *Taxodium–Ulmus* forests with *Glyptostrobus, Nyssa, Alnus*, and Myricaceae. Riparian assemblage of *Nelumbo, Potomogeton*, Nymphaceae, *Sparganium, Arundo, Strelitzia, Aponogeton*, and Cyperaceae	Rhinocerotids (*Protaceratherium*). Broad-leaved and coniferous–broad-leaved polydominant forests of the Turgai type with Juglandaceae, *Fagus, Liquidambar, Acer*, and abundant Rosaceae, Anacardiaceae, and *Corylus* in underwood	Indricotheriinae (*Paraceratherium*). Mixed polydominant forests with *Fagus, Quercus, Zelkova, Carya*, and thick underwood. Semienclosed *Pinus–Quercus* forests in elevated areas
				Indricotherium			⊖			

(Contd.)

Age, Ma	Series	Subseries	Stage	Types of mammalian fauna	Climate		Transgressive ⊕ and regressive ⊖ phases	Marshy brushwood and riparian forests	Forests in flat lowlands	Vegetation of elevated plains, thin and savanna-like forests, brushwood (open and semiclosed vegetation)
29.6	Oligocene	Lower	Rupelian	*Brontotherium*	Semihumid	Moderately warm, sometimes subtropical		Amynodonts (*Cadurcodon*) and suids (*Anthracotherium*). *Taxodium–Nyssa* and *Taxodium–Ulmus* forests with *Sassafras, Alnus, Populus*, Myricaceae, *Arundo*, Nelumbaceae, Nymphaceae, *Zingeberopsis, Strelitzia*, and *Ceratopyllum*	Creodonts (*Hyaenodon*), chalicotheres (*Schizotherium*), and hyracodonts (*Prohyracodon*). *Taxodium*–broad-leaved forests with *Liquidambar, Platanus, Ulmus, Ailanthus, Eucomia, Acer, Tilia, Zelkova, Castanea, Quercus, Spiraea* brushwood, Leguminosae, *Alnus, Rhus*, and *Comptonia*	Tapiroids (*Colodon*), hyracodonts (*Ardynia, Tenizia, Indricotherium*), suids (*Enteledon*), and ruminants (*Palaeohypsodon*). Thin *Quercus*–Lauracea–Leguminosae forests with shibliak and other bush formations (*Rhus, Palibinia*, Myricaceae, and Myrtaceae)
30						Subtropical, approaching moderately warm	⊕	I. Brontotheres (*Embolotherium*) and amynodonts (*Cadurcodon, Gigantamynodon*); II. Amynodonts (*Sharamynodon, Gigantamynodon, Syanodon*) and suids (Anthracotheriidae). *Hibiscus–Populus–Ulmus* forests with Juglandaceae, *Alnus, Nyssa, Magnolia*, Myricaceae, and Leguminosae. Riparian brushwood with *Arundo, Zingiberopsis, Nelumbo*, and aquatic ferns (*Azolla, Salvinia*)	I. Creodonts (*Hyaenodon*), chalicotheres (*Schizotherium*), and ruminants (Tragulidae); II. Brontotherioids (*Rhinotitan*), chalicotheres (*Schizotherium*), creodonts (*Propterodon*), and ruminants (*Archaeomeryx*). *Taxodium–Sequoia*–broad-leaved forests with *Castanopsis, Lindera, Liquidambar*, Hamamelidaceae, *Sassafras, Magnolia*, and *Rhus–Ilex* brushwood	I. Tapiroids (*Colodon*), hyracodonts (*Ardynia*), Indricotheriinae (*Urtynotherium*), and suids (*Entelodon*). II. Indricotheriinae (*Djuxia*), tapiroids (*Deperetella, Colodon*), and suids (*Eoentelodon*). *Pinus–Dryophyllum* and *Quercus*–Lauraceae forests, thin savanna-type forests, and brushwood with Araliaceae, Anacardiaceae, and *Ilex*
35							⊖			
36.4	Eocene	Upper	Priabonian		Semihumid, and semiarid in the second half	Subtropical (periodically humid)	⊕			
40		Middle	Bartonian	Dinocerata–Condylarthra–Tapiridae	Semiarid–semihumid	Seasonal subtropical with hot summer period	⊖	Dinocerata (*Gobiatherium*), Condylarthra (Mesonychidae), and Tapiridae (*Breviodon*). *Hibiscus–Populus* forests with bushes of Leguminosae, *Indesia, Ungnadia, Cupanites*; brushwood of *Bambusa, Arundo*, Nelumbaceae, and Nymphaceae	Tapirids (Isectolophidae), brontotheres (*Protitan*), and Condylarthra (*Paraphenocodus*). *Dryophyllum* forests with *Castanopsis, Zelkova, Celtis, Liquidambar, Ficus*, Leguminosae, *Cinnamonum*, and *Magnolia*	Tapirids (*Lophialetes*) and rhinocerotids (*Forstercooperia*). *Sequoia–Quercus*–Lauraceae forests with Leguminosae and *Ailanthus*. Brushwood of Proteaceae and Ericaceae; bush–herbaceous assemblages of *Ephedra*, Chenopodiaceae, *Nitraria*, and Leguminosae
42.1							⊕			

ground, the moderately thermophilic flora gradually replaced (during a period of several m.y.) the autochthonous subtropical vegetation, adding deciduous (Turgai) elements to its composition. The invasion of thermophilic, and then mesophilous components was promoted by the marine basin regression from the territories of the West Siberia and Turgai depression, and subsequently from the regions adjacent to the eastern Urals. The sea left behind vast coastal plains, where the *Taxodium–Nyssa–Alnus* forests were under development. The more open landscapes and saline soils were colonized by Myricaceae, *Ephedra*, and Chenopodiaceae. The *Alnus–Populus–Ulmus* communities occupied river valleys of a newly formed hydrographic network; they included such forms as *Liquidambar, Acer, Rhus*, and different Juglandaceae (*Carya, Pterocarya, Juglans,* and *Engelhardtia*). Laurofoliate forms and microfoliate oaks still retained their significance, primarily in elevated plains. Forests of mountainous regions (Altai) included conifer and broad-leaved taxa (*Pinus, Metasequoia, Thuja, Sciadopitys, Tsuga*, Betulaceae, and Juglandaceae). This distribution of taxa in the early Oligocene plant communities is inferred from the taphocoenoses of the Turgai depression. The taxonomy of local floras from a single stratigraphic level, where they associate with abundant vertebrate remains of the so-called *Indricotherium* assemblage, is diverse. Lentils of coarse alluvium usually bear small leaves of sclerophylls derived from areas elevated over the valleys which were better desiccated at that time. Sclerophylls formed thin bush-arboreal forests, which included Lauraceae, Fagaceae, *Rhus,* Ericaceae, Leguminosae, and other taxa. Plants, buried in clay deposits of oxbow lakes and flood plains, are also diverse in composition. Typical representatives of the respective riparian forests were *Populus, Alnus, Ulmus*, and Juglandaceae associated with Taxodiaceae, herbs, and aquatic plants (*Typha, Arundo, Strelitzia, Zingiberopsis, Nelumbo, Salvinia*, and others). Sometimes, these two groups of taphocoenosis are insignificantly separated in sections by sedimentary lenses bearing remains of rodents, carnivores, and large rhinocerotids also typical of different biotopes. Burials of this type have been reported from the Manyske-Suek and Shintuzsai localities.

In more arid regions of Kazakhstan and Central Asia, the stunted savanna-type forests consisted of laurofoliate forms, Anacardiaceae, and Leguminosae. Their sclerophyllous shrubs included *Rhus turcmenica, Palibinia*, and xeromorphic *Myrica*. Open areas, including those on the river banks, were favorable for Gramineae, Chenopodiaceae, Compositae, and Cyperaceae. During the first half of Oligocene, the climate in the southern and central regions of Kazakhstan was still seasonal, with hot and dry summers. The inferred character of vegetation cover was typical of the period, when the Brontotheriinae and Indricotheriinae faunas were at the final and initial stages of evolution, respectively.

This period is marked by the extinction of some perissodactyls (brontotheres) adapted to humid environments, and by the abruptly decreased diversity of tapirids. The brontoterioid–tapiroid communities were replaced by taxa of swamp rhinoceros (Amynodontidae) and hyracodonts (*Ardynia*), which included more progressive forms of similar adaptation ability. The swamp rhinoceros, who fed on aquatic and riparian plants, had reduced incisors, large canine teeth, and an increased molar row (masticators) at the expense of premolar teeth. The dentition of hyracodonts (*Ardynia*) was peculiar because of the wide-crowned incisors and certain similarity between the canine and frontal teeth. These animals also had more hypsodont teeth in the cheek row and longer limbs, especially in distal parts. Giant rhinoceros (*Indricotherium*) lost many elements of the incisor–canine system, with the exception of two nipping teeth, which resembled small tusks. However, the brachydont character of their cheek teeth clearly indicates that these animals fed on soft vegetation, presumably on leaves and shoots of tall trees. This group of animals coexisted with early representatives of Ruminantia (Tragulidae). The dentition of Oligocene perissodactyls demonstrates a wide spectrum of adaptation to more sclerophyllous vegetal food. The main features of their dentition were as follows: an enlarged area of incisor–canine teeth; molarization of premolars; higher crowns of cheek teeth; a more complex relief of these crowns ("crochet" and "anticrochet" system) responsible for their rigidity; the appearance of dental cementum on the crowns; and a larger total area of molars (hypsodonty of M_2^2).

At the mid-Oligocene (Solenovo) time, when the epicontinental marine basin was transformed into that of the brackish-water type, the climate in the southern Russian platform and Kazakhstan was more distinctly seasonal than before. This gave rise to the partial redistribution of dominant forms in zonal forests. Pinaceae became more abundant than Taxodiaceae, and Fagaceae prevailed over Betulaceae and Juglandaceae. At the middle-late Oligocene boundary, the Zaisan depression and bordering areas of Kazakhstan were mostly occupied by coniferous to broad-leaved forests with abundant lianas. The generic spectrum of these forests was similar to that typical now of the woodlands of northeastern China, Korea, and southern Primor'e. Being extremely diverse, the arboreal species formed a series of individual botanical formations. In the coastal flatland with numerous rivers (Turgai and Aral regions), the main silva-forming taxa were Taxodiaceae and Nyssaceae (the latter in more marshy lands). In the polydominant forests of elevated areas, the principal forms were alternately represented by *Tsuga, Cedrus, Pseudotsuga, Sciadopitys, Picea, Betula*, or *Carya*. Thinner forests, especially in the Turgai region, consisted of *Pistacea, Rhus, Ilex*, Leguminosae, *Ziziphus, Paliurus*, and *Celtis*. They alternated in space with pine forests and oak woods, which still retained some Lau-

raceae. The underwood included diverse bush and grass taxa, such as Leguminosae, Umbelliferae, forbs, Liliaceae, Anacardiaceae, Celastraceae, and Chenopodiaceae (palynospectra from the well-known Chelkar-Teniz locality display up to 30% of pollen of the last taxon). Between the forests, there were also open areas free of arboreal plants and occupied only by shrubs and herbs. These areas were likely a refugium of large rhinocerotids and other mammals, including abundant rodents.

The riparian vegetal formations were represented by marshy forests in river valleys, where they included *Alnus, Nyssa, Taxodium, Glyptostrobus, Populus*, and *Ulmus*. These forests prevailed in deltaic areas of large rivers, and sometimes were of the distinct bidominant type, when the principal *Populus–Ulmus* assemblage associated with *Pterocarya, Phellodendron, Fraxinus, Alnus, Betula, Liqudambar*, and *Acer*. Assemblages of Gramineae, *Sparganium, Typha*, and *Arundo* were widespread in coastal zones of lakes, oxbows, and river canals. These riparian and aquatic assemblages still included *Salvinia, Nymphaceae, Nelumbo, Potomogeton, Trapa*, and *Strelitzia*, and provided favorable environments for turtles (*Trionyx*). All of these features characterize the zonal and intrazonal vegetation, which preserved relics of the Eocene subtropical vegetation at the initial mid-Oligocene stage of evolution and was designated by A.N. Krishtofovich as the Turgai flora.

Late Oligocene

The activated erosion of the late Oligocene–early Aquitanian interval was a response to the further regression of the marine basin. Due to subsequent cooling, conifers and mesophilous amentiferous *Carpinus, Ostrya, Betula*, and Juglandaceae became of prime importance among the dominants of woody vegetation. The birch–walnut forests replaced the *Taxodium–Glyptostrobus* assemblages in river valleys, because the lowered erosion base made soils better drained. As before, coenoses of Gramineae prevailed along the river banks.

In Kazakhstan, the boundary between the arid and humid zones of the late Oligocene is confined to the transition from red clayey deposits of the Bet-Pak-Dala and Saryin formations to greenish-gray sediments of the Chelkar-Nura (Turgai region) and Chilikta (Aral region) formations. Crossing the Turgai upper courses westward of Aral'sk, this boundary is traceable in the south near the eastern scarps of the Chelkar-Teniz Lake (Atambas-Chink), and then extends into the southeastern Aral region. It is remarkable that most burials of the *Indricotherium* fauna are concentrated within this transition (ecotone) zone.

Over the area of red deposits, the climate was of a clearly seasonal type, and the mosaic vegetation here included a wide spectrum of woody formations. The somewhat greater humidity of the mid-late Oligocene climate in the south and east of Kazakhstan was responsible for the lesser differentiation of vegetal cover. The mesophilous coniferous and broad-leaved forests of that time expanded over Central Asia up to northwestern China.

Consequently, the burials of mammalian remains in the upper Oligocene deposits are less abundant, and their taxonomic diversity is lower. For instance, we know dozens of localities with *Indricotherium* remains, while burials of its direct progeny *Paraceratherium*, the largest form of rhinocerotids, are extremely rare. The late Oligocene was a prosperous time for the mesophilous deciduous flora of the Turgai type, whose very rich assemblages were recovered in the famous Ashutas and many other localities of the Turgai and Aral regions, where they are confined to the upper horizons of the Chilikta and Chagrai formations, and to other coeval deposits.

Early Miocene

At the commencement of Neogene, namely at the accumulation time of the Aral Formation, the climate was again more seasonal and dry. In response, the pine-oak forests gradually replaced Fagaceae (in elevated areas) and birch-walnut forests (in depressions). Landscapes also changed, and vast areas were occupied by lakes. In the words of V.V. Lavrov, it was the "great stage of lakes" in Asia. Shrubs and herbs invaded the vegetal cover, where woody forms were losing their significance. Occurrences of leaf flora in Kazakhstan are rare, and we used palynological data to depict the plant evolution. In response to aridization, mesophilous elements of the Turgai type decreased in abundance and partially migrated into azonal floral groupings. In zonal complexes, they were replaced by mesoxerophilic taxa, whose descendants still exist in the east Mediterranean region. The Kushuk flora is surprisingly lacking in conifers, and includes only microfoliate arboreal and dumetosous taxa (*Castanea, Juglans, Pterocarya, Ulmus*, and *Zelkova*). Palynocomplexes from the lacustrine facies of the Aral Formation are different from those typical of the Oligocene time. They correspond to a new type of zonal vegetation rich in herbs and shrubs. Subordinate arboreal taxa are represented in this case by *Pinus, Picea, Quercus, Tilia, Betula, Alnus, Salix*, and *Ulmus*. In addition, *Castanea, Celtis, Carpinus, Acer, Zelkova*, and *Pterocarya* occur in southern assemblages. Judging from the composition of the herbs, they were ancestors of later steppe assemblages, and included Chenopodiaceae, Leguminosae, Caryophillaceae, Polygonaceae, Umbelliferae, Ericaceae, and sometimes, Gramineae. *Potomogeton, Sparganium, Myriophyllum*, and *Nelumbo* grew in water basins. The *Ephedra* genus was diverse. Some communities included psammophytes (*Calligonum* and others). In more northern areas, e.g., in southern West Siberia, forest vegetation was also losing its former "Turgai features." Progressing new communities consisted of species currently characteristic of Siberia.

However, some elements of the Turgai flora (*Juglans, Acer, Fraxinus*, and others) migrated into forests of the bush-steppe belt of the Tien Shan, Xinjiang, and Alai. They existed there throughout the Neogene.

The earliest Miocene was the extinction time of the *Paraceratherium* ungulate complex of Central Asia. Other groups of mammals also experienced significant changes. The place of giant rhinocerotids was occupied by ecologically similar mastodonts (*Gomphotherium*), swamp rhinoceroses (Amynodontidae) were replaced by true rhinoceroses (*Brachypotherium*), and mobile hyracodonts (*Ardynia* and *Allocerops*) by larger *Aceratherium* forms resembling their ancestors in adaptation features. It was also the formation time of the basic taxa of artiodactyl Ruminantia, which were at the commencement of their prosperity.

CONCLUSION

The detected dynamics of evolutionary changes in vegetal and mammalian communities is shown in the table.

ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research, project no. 93-05-09087.

Reviewer A.G. Ponomarenko

REFERENCES

Makulbekov, N.M., *Eotsenovaya flora Severnogo Kazakhstana* (Eocene Flora of Northern Kazakhstan), Alma-Ata: Nauka, 1972.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 378–391. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 69–82.
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Zoogeography of the Late Oligocene Basins of Western Eurasia Based on Bivalve Mollusks

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Received January 21, 1994

Abstract—The main difference between the distribution of Chattian and previous molluscan assembalges in Western Eurasia was the increase in the degree of similarity of all the assemblages due to smooth climatic barriers and the extension of genera and species distribution areas. The north European and Mediterranean provinces continued to exist, but a wide transitional zone appeared between them. As earlier, the German and East Paratethyan subprovinces may be distinguished.

Key words: *biogeography, Paleogene, bivalve mollusks, Tethys, Paratethys.*

The evolution of the late Oligocene marine benthic fauna proceeded under general sea regression and initial activization of the Alpine orogeny. In the West European basins, the sea retreated, and mainly continental deposits accumulated. The area occupied by marine basins also decreased in southern Europe. However, at the beginning of the Chattian, the Paratethys connected with the Norwegian-Greenland sea, and subsequently with the Tethys. Sedimentation and fauna evolution proceeded here in mild climatic environments without distinct barriers against the background of positive tectonic movements and denudation, which provided a supply of sandy material into the basins. The characteristics of the material and methods were described in an earlier publication (Popov, 1994). The revision of Oligocene mollusks (Popov *et al.*, 1993) allowed a single list of mollusks from the considered regions to be compiled (Table 2) and species comparison (Table 3) to be performed. To shorten the species list, species that are known from only one region are marked in Table 2 by numbers, which designate the number of genus (subgenus) characteristic species in the considered region.

CHARACTERISTICS OF THE ASSEMBLAGES

In North Africa, the Chattian molluscan complexes are, evidently, unknown thus far.

Southern Europe. The only representative Chattian molluscan complex was reported from northern Italy (Venzo, 1937; Accordi, 1955). According to these researchers, more than 100 species of 64 genera and subgenera are present here (Table 2; Fig. 1), among which are such tropical and subtropical genera as *Trisidos, Spondylus, Florimetis, Crassatella* s.s., *Cardium* s.s., *Vepricardium, Trachycardium*, and *Venus* s.s. Ten genera and approximately 60% of species mainly representing these most thermophilic taxa are unknown outside southern Europe (Group I in the Table 2). Yet, the significant share of this complex belonged to species of wide geographic distribution (22%). Even more diverse were the species in common with those from the central European basins. The share of species of north European origin in the Tethyan fauna continued to increase: *Astarte concentrica, Nemocardium cingulatum, Arctica rotundata, Gallista reussi,* and *Gari angusta* can be mentioned to exemplify the new probable migrants from the northwest European seas.

Transcaucasia. In the territory of Armenia, Chattian marine fauna is unknown because the marine basin ceased to exist there at that time. In Georgia, the Chattian molluscan assemblage is not very representative, being composed of 33 species. It includes such warm-water subtropical forms as *Arca biangula, Barbatia (Obliquarca), Isognomon, Arcopagia*, and *Corbula (Caryocorbula)*. However, these genera and subgenera are represented by widely distributed species and lack typical Tethyan forms.

Central Europe. The Chattian (upper Kiscellian and lower Egerian) deposits of Central Europe (in Hungary, Romania, and Austria) incorporate rich representative assemblages of bivalve mollusks, which are important with respect to their zoogeography: this fauna shows great similarity to mollusks from both southern and northwestern Europe. Assemblages of Central Europe are sufficiently close to each other (K_p = 0.60–0.72, see Table 1), and include various thermophilic, subtropical and even tropical genera (*Pinctada, Isognomon, Crassatella, Linga, Megacardita, Cardium* s.s., and *Vepricardium*). Among the thermophilic mollusks, 19 species are in common with the Mediterranean forms (Table 2) and unknown from more northern assemblages. They include species characteristic of the Tethyan fauna since the Eocene, such as *Pecten arcua-*

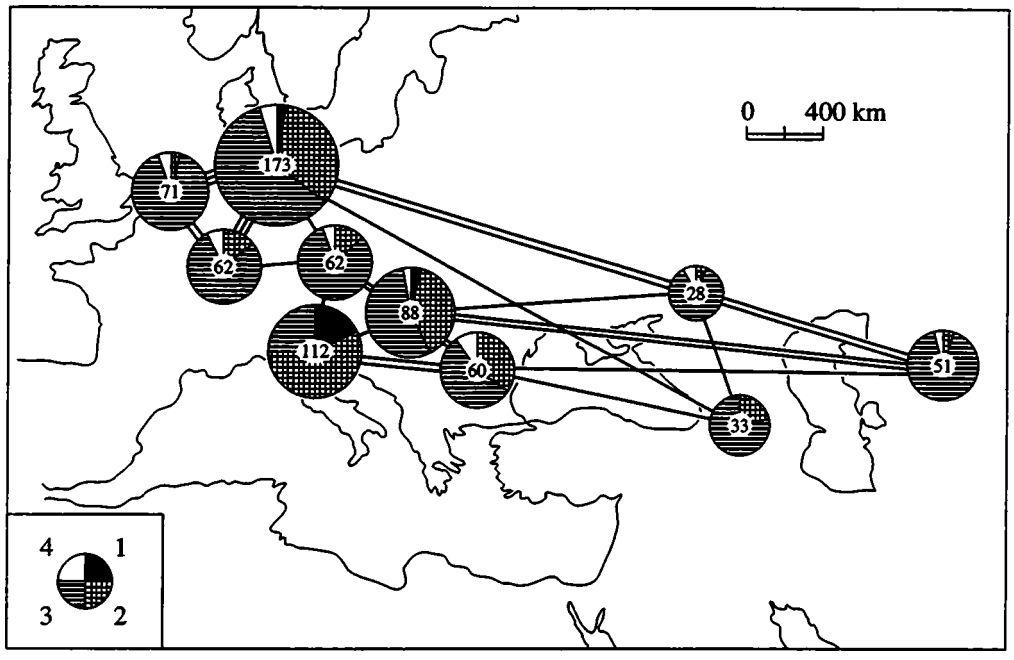

Fig. 1. Circle diagrams showing the diversity and thermophilic affinities of the Chattian bivalve assemblages. The circle sizes are proportional to the species number in the center of the circle.

1, Relative share of species of tropical genera; 2, relative share of species of tropical and subtropical genera; 3, relative share of widespread species; 4, relative share of predominantly boreal species. Lines connecting the circles indicate the degree of similarity between assemblages.

tus, Picnodonte brongniarti, and *Crassatella carcarensis*, as well as *Pecten (Flabellipecten) burdigalensis, Megacardita arduini, Lutraria oblonga, Peronaea planata, Venus multilamella*, and *Pholadomya alpina*, which, probably, first appeared in the Mediterranean, and many of which also became widely distributed in the Miocene in northern assemblages. At the same time, the share of species that render the considered assemblages as similar to the northern faunas is also significant. These species include such characteristic forms as *Chlamys picta, Spondylus tenuispina, Gonimyrtea schloenbachi, Scalaricardita orbicularis, Astarte gracillis, Callista beyrichi*, and others. It should be remembered that throughout the late Kis-

Table 1. Comparison of the species composition of the Chattian bivalves from western Eurasia. Preston's coefficients

Region	1	2	3	4	5	6	7	8	9	10
1. Italy	**112**	0.69	0.72	0.76	0.80	0.90	0.90	0.89	0.90	0.87
2. Romania	22	**60**	0.69	0.71	0.71	0.82	0.86	0.84	0.86	0.85
3. Hungary	34	28	**88**	0.60	0.61	0.72	0.73	0.85	0.80	0.81
4. Austria	28	24	40	**62**	0.72	0.76	0.81	0.84	0.87	0.85
5. Bavaria	20	22	34	20	**62**	0.61	0.56	0.79	0.74	0.73
6. Belgium + the Netherlands	12	15	28	23	31	**71**	0.49	0.80	0.70	0.68
7. N. Germany + Denmark	20	17	39	26	45	58	**173**	0.73	0.69	0.75
8. Georgia	9	9	10	10	11	12	20	**33**	0.85	0.91
9. Black Sea region, Volga-Don region	7	7	11	8	12	15	20	6	**28**	0.45
10. Transcaspian region	11	11	16	12	17	23	27	6	22	**51**
	Number of species in common									Number of species

Note: Semibold numbers placed diagonally indicate species in the assemblages; high Preston's coefficient values are hatched; the coefficents showing the closest connection are doubly hatched (for treshhold values see Popov *et al.*, 1993; Popov, 1994).

Table 2. Distribution of bivalve taxa in the late Oligocene basins of western Eurasia (figures show the number of species recorded in the region)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
I. Genera and species of the Tethyan distribution										
Barbatia (Barbatia)	3			?						
Anadara	2									
Trisidos	2									
Glycymeris	4									
Chlamys (Aequipecten)	8									
Spondylus	2									
Lucina	4									
Megaxinus	3									
Anodontia fragilis (Phil.)	+									
Florimetis lamellosa (Lmk.)	+									
Chama	3									
Cardita	3									
Cardiocardita	3									
Carditamera (Lazariella) desioi (Venzo)	+									
Astarte aff. *solida* Desh.	+									
Crassatella oligocenica Rov.	+									
Trachycardium	2									
Loxocardium distinguendum (May.)	+									
Plagiocardium fallax (Micht.)	+									
Nemocardium (Nemocardium)	3									
N. (Habecardium) peracutum (Rov.)	+									
Lutraria latissima Desh.	+									
Tellina granconensis Opp.	+									
Arctica stefaninii (Venzo)	+									
Glossus justinensis (May.)	+									
Venus (Venus)	3									
V. (Ventricoloidea) praecursor (May.)	+									
Clausinella	3									
Chione delbosi (Micht.)	+									
Tapes destefanii (Canest.)	+									
Dosinia basteroti (Agas.)	+									
Callista lesbarritzensis (Coss.)	+									
II. Species in common between the Tethys and central Europe										
Pecten (Pecten) arcuatus Br.	+		+							
P. (Flabellipecten) burdigalensis Lmk.	+	+	+							
Chlamys (Aequipecten) northamptoni (Mich.)	+		+							

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Pycnodonte brongniarti Br.	+		+							
Megaxinus deperdita (Micht.)	+		+							
Megacardita arduini (Br.)	+		+							
Crassatella carcarensis Mich.	+		+	+						
Cardium bukkianum Roth.	+		+							
Vepricardium egerense (Roth.)	+		+							
Nemocardium (Habecardium) semilineatum (Koen.)	+		+	+						
Lutraria angusta precedens Venzo	+	+								
L. oblonga (Chemn.)	+	+	+	cf.						
Peronaea planata (L.)	+		+							
Macoma elliptica Br.	+		+							
Venus (Ventricoloidea) multilamella Lmk.	+	+	+							
Nitidavenus subarata (Sandb.)			+							
Corbula (Caryocorbula) carinata Duj.	+	+	+							
Pholadomya (Pholadomya) alpina (Math.)	+			+						
Thracia bellardi (Pict.)	+	+	+							
Clavagella oblita Mich.	+		+	+						
III. Genera and species of the central European distribution										
Nucula (Nucula) mayeri Hörn.			+							
Nuculana (Costatoleda) psammobiaeformis (T. Roth.)		+	+							
N. (Saccella)			2							
"Nuculana" korobkovi			+							
Saturnia raulini Coss., Peyr.			+							
Anadara gumbeli May.			+	+						
Arcopsis			1	1						
Trinacria oligocenica Baldi			+							
Glycymeris latiradiata Sandb.			+	+						
Mytilus aquitanicus May.		+	+	+	+					
M. galloprovincialis dilatatus Phil.		+								
Modiolus taurinensis (Bon.)				+						
Chlamys			1	1						
Propeamussium			2							
Palliolum (Palliolum) incomporabile Risso			+	+						
P. (Delectopecten) unguiculus (May.)			+							
Spondylus crenellaeformis Nosz.			+							
Acesta miocenica Sism.			+							
Kiscellima cancellata (Hofm.)			+							
Limaria langniana Sacco			+							

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Cubitostrea frondosa (de Sir.)		cf.		+						
Crassostrea aginensis (Tour.)		+		+						
Lucina sp.			+							
Linga oligobligua (Sacco)		+								
L. agassisi (Mich.)			+							
Gonimyrtea intercalata (Hölzl)		+								
Saxolucina incrassata (Orb.)		cf.		+						
Saxolucina			2							
Divalinga regaultiana (Desh.)			+	+						
Loripes dentatus (Defr.)		+								
Thyasira vara Kor.			+							
Glans ruginosa Cos., Peyr.			+							
Cardiocardita monilifera Duj.			+							
Crassatina bosqueti Koen.			+							
"*Cardium*" *deminutum* Hölzl		+								
Laevicardium praesandbergeri Hölzl		+								
Acanthocardia bojora (May.)		+	+							
Cerastoderma		2								
Lutraria		3		1						
Solen parvus Hölzl			+							cf
Cultellus budensis Baldi			+							
Aquincosiliqua hermati (Nosz.)			+							
Capsa lacunosa Chemn.			+							
Arcopagia subelegans Orb.			+							
Gari (*Gobreus*)		4								
Abra budensis Hofm.			+							
Congeria		3								
Polymesoda (*Batissa*)		2								
Pelecyora (*Cordiopsis*) *boehmi* (Hölzl)		+								
Corbula applanata				+						
Lentidium		2								
Pholadomya (*Pholadomya*)			2							
Cuspidaria			3							
Cardiomya noszkyi Baldi			+							
Pecchiolia argentea Mar.			+							
IV. Widespread species										
Arca biangula Lmk.	+		+	+			+	+		
Glycymeris lunulata (Nyst.)	+		+	cf	+	+	+	+	+	+

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Isognomon maxillatus (Lmk.)	+		+	+			+	cf		
Anomia striata Br.	+					+				
Crassostrea cyathula (Lmk.)	+	+	+	+	+		+			
Licinoma borealis (L.)	+		+	+	+		+	+	+	cf
Astarte concentrica Goldf.	+				+		+			
A. kickxii Nyst.	+				+		+			
Parvicardium michelottii (Desh.)	+							+		
Nemocardium (*Habecardium*) *cingulatum* (Goldf.)	+			+	+		+			+
N.(*H.*) *tenuisulcatum* (Nyst.)	+	+	+	+	+	+	+			
Gari (*Psammotaena*) *angusta* (Phil.)	+	+	+		+	+	+			+
Azorinus antiquatus Pult.	+		+							+
Arctica rotundata (Ag.)	+	+	+	+	+	+	+		+	+
Glossus subtransversus (Orb.)	+	+	+	+	+	+	+	+	+	+
Polymesoda brongniarti (Bast.)	+	+	+					+		
Callista splendida (Mer.)	+	+	+	+	+	+	+			
C. reussi (Spey.)	+				+	+	+	+		
Pelecyora (*Cordiopsis*) *westendorpi* (Nyst.)	+	+	+	+	+	+	+	+		+
Corbula (*Varicorbula*) *gibba* (Ol.)	+	+	+		+	+	+	+	+	+
Panopea intermedia (Sow.)	+	+		+	+	+	+	cf	+	+
P. meynardi (Desh.)	+	+	+	+	+	+	+			cf
Pholadomya (*Bucardiomya*) *pushi* Goldf.	+	+	+	+	+	+	+			
Thracia ventricosa Phil.	+	+	+	+	+	+	+			
V. Species absent from the Tethyan basin										
Nucula (*Nucula*) *schmidti* Glib.			+		+	+	+			
N.(*N.*) *nucleus* L.			+	cf		+				
N. (*Lamellinucula*) *comta* Goldf.			+	+	+	+	+	cf	+	+
Leionucula burdigalica (Cos. Peyr.)			+							cf
L. laevigata (Sow.)				+	+	+	+		+	+
Yoldia glaberrima (Münst.)			+	+	+		+			
Limopsis (*Pectunculina*) *retifera* Semp.			+	+			+			
Glycymeris obovata (Lmk.)				+		+	+			
Mytilus acutirostris Sandb.		+				?		cf		+
Modiolus dunkeri (Sp., Koen.)			+				+			+
Musculus philippii Wolf.			+				+			
Arcoperna micans (Br.)			+	+	+	+	+			
Atrina pectinata (L.)				cf			+			
Pinctada phalaenacea (Lmk.)			+	cf	+		+			
Pteria hirundo L.			+				+			

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Chlamys (*Aequipecten*) *picta* Goldf.			+	+			+		+	+
Palliolum decussatum (Münst.)			+	+		+	+			
Spondylus tenuispina Sandb.				cf			+			
Anomia ephippium (L.)			+	+			+			
Pycnodonte callifera Lmk.		+	+	+	+	+	+	+		+
Lucinoma barrandei (May.)				+						+
Gonimyrtea schloenbachi (Koen.)			+			+	+			
Diplodonta rotundata Mont.			+	+		+				
D. trigonula (Br.)		+						+		
Felaniella fragilis (Sandb.)			+		+	+	+			
Scalaricardita orbicularis (Sow.)			+		+	+	+		+	cf
Astarte (*Astarte*) *gracilis* (Münst.)			+		+	+	+		+	cf
Parvicardium neglectum (Hölzl)		+	+		+			+		
Ensis hausmanni (Goldf.)			+		+	+	+			+
Siliqua nysti Desh.			+				+		+	
Angulus (*Angulus*) *nysti* Desh.		+	+	+	+	+	+		+	+
A. (*Moerella*) *posterus* Beyr.			+		+	+	+			+
Solecurtus basteroti Des Monl.			+		+		+			
Callista (*Costacallista*) *beyrichi* (Semp.)		+	+	+	+	+	+	+		
Dosiniopsis sublaevigata (Nyst)			+			+				
Janschinella vinogrodskii (Merkl.)		+							+	
Corbula (*Corbula*) *basteroti* Desh.			+		+	+				
C. (*Lenticorbula*) *helmerseni* (Mich.)		+								+
Hiatella arctica (L.)			+			+	+			
Teredo sp.				+			+			
Cuspidaria clava Beyr.			+	+	+		+			
C. cuspidata (Ol.)			+	+		+				+
VI. Species in common between the central Europe and eastern Paratethys										
Nucula (*Lamellinucula*)							2			
Nuculana (*Saccella*) *westendorpi* (Nyst)					+		+			
N. (*S.*) *gracilis* (Desh.)						+	+	+		+
N. (*Jupiteria*) *pygmaea* (Münst.)					+	+	+	+		+
N. (*J.*) *nana* Koen.										+
Yoldia strucki (Wiech.)							+			
Barbatia (Barbatia)							2			
B. (*Obliquarca*) *modioliformis* (Desh.)								+		
B. (*Cucullaearca*) *scabrosa postera* (Sandb.)							+			
Acar aff. *dentiens* (Coss., Peyr.)							+			

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Bathyarca bellula (Wiech.)							+			
Bathyarca sp.										+
Arcopsis pretiosa (Desh.)							+	cf		
Limopsis (*Limopsis*) *aurita* (Br.)							+			
L. (*Pectunculina*) *goldfussi* Nyst							+	+		
Nucinella dobergensis (Lien.)							+			
Glycymeris agnosta Jan.							+			
G. philippi (Desh.)						+	+			
Mytilus chatticus Gor.							+			
Septifer gramiliferus (Spey.)					+		+			
Modiolula pygmaea (Phil.)							+			
Crenella beyrichi (Spey.)							+			
Gregariella sp.							+			
Modiolus penndorfi (Görg.)							+			
M. katscharavai (Char.)								cf		
Musculus stenbergensis And.							+			
M. tamdiensis Mir.										+
Lithophaga caeca Str.							+			
Isognomon heberti (Coss., Lamb.)							+			
Chlamys (*Hilberia*) *bifida* (Münst.)						+	+		+	+
Ch. (*Hilberia*)							4			
Ch. (*H.*) *hoeninghausi* (Defr.)						+	+			+
Propeamussium (*Parvamussium*) *pygmaeum* (Münst.)							+			
Palliolum (*Palliolum*)							3			
P. (*P.*) *hausmanni* (Goldf.)					+		+			
P. (*P.*) *striatocostata* (Goldf.)						+	+			
P. (*Delectopecten*) *simile* (Lask.)						+	+		+	+
Plicatula casselensis (Jan.)							+			
Anomia bullata Phil.					+		+			
A. asperella Phil.							+			
A. goldfussi Desh.						+	+	cf	cf	
Placuna philippi (Sp.)						+	+			
Pododesmus (*Heteranomia*) *squamula* (L.)							+			
Limaria sandbergeri (Desh.)							+			
Limatula							2			
Limea (*Notolimea*) *nysti* (Spey.)							+			
Ostrea sp.								+		
"*O.*" *nettelbladti* Weich.							+			

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Lucinoma praecedens (Koen.)					+	+				
Parvilucina thierensi (Heb.)							+			
Megaxinus (*Gibbolucina*) *gibbosula* Lmk.										cf
Saxolucina omaliusi (Desh.)										cf
Codalucina tenuistria (Heb.)								cf		
Ctena squamosa (Lmk.)							+			
Paralucinella undulata (Lmk.)						+	+			
Thyasira						1	1			1
Diplodonta speyeri							+			
Felania schmidti (Gorg.)							+			
Spaniorinus						2	1			
Pteromeris laevigata (Spey.)						+				
Scalaricardita							2			
S. depressa (Koen.)						+	+			
Carditopsis clara (Koen.)								+		
Astarte (*Astarte*) *henkeliusiana* Nyst					+		+			
A. (*A.*) *pygmaea* Münst.						+	+			
A. (*A.*) *trapeziformis* (Spey.)							+			
A. (*A.*) *goldfussi* Hind.					+	+	+			
A. (*Laevastarte*) *angulata* (Lehm.)							+			
A. (*Digitariopsis*) *propinqua* (Goldf.)						+	+			
A. (*D.*) *falsopropinqua* (Jans.)							+			
A. (*Digitaria*) *koeneni* (Spey.)						+	+			
Goodallia laevigata (Spey.)						+	+			
Grotrania lunularis (Phil.)							+			
G. semicostata Spey.						+	+			
Crassatina (*Chattonia*) *astarteiformis* Nyst							+			
Plagiocardium abundans (Liv.)									+	+
Parvicardium kochi (Semp.)					+		+			
P. lamberti (Cossm.)								+		
P. scobinula (Desh.)						+	+			
P. nikolaevae Merk.									+	
Cerastoderma mutabile Pop.									+	+
C. karaschokense Merk.								cf	+	+
C. prigorovskii Bog.									+	+
C. chamidi (Bel.)										+
Nemocardium (*Habecardium*) *excomatulum* (Gl., Poel)					+	+	+	+		
Mactra sp.								+		

Table 2. (Contd.)

Genera and species \ Geographic distribution	Italy	Romania	Hungary	Austria	Bavaria	Belgium + the Netherlands	Northern Germany + Denmark	Georgia	Black Sea region, Volga-Don region	Transcaspian region
	1	2	3	4	5	6	7	8	9	10
Spisula (*Spisula*) *subtruncata* (Da C.)					+	+	+			
S. (*Pseudoxyperas*) *westphalica* (Gorg.)							+			
Cultellus roemeri Koen.				cf		+	+			+
Siliqua							1			2
"*Tellina*" *pusilla* Phil.					+	+				
T. (*Elliptotellina*) *longiuscula* Koen.							+			
Angulus (*Moerella*) *casselensis* (Jan.)							+			
Peronae aquitanica (May.)							+			
Arcopagia					1		1	1		
Abra bosqueti (Semp.)					+	+	+			
Solecurtus philippi (Spey.)						+	+			
Pygocardia cyprinoides (Sandb.)						+	+		+	+
Trapezium usturtense (Il.)										+
Alveinus nitida (Reuss.)							+			
Miocardiopsis supraoligocenica Jan.							+			
Polymesoda convexa (Bron.)					+		+			
Venus (*Ventricoloidea*) *koeneni* Lien.							+			
Callista							3			
Pitar (*Calpitaria*) *andersoni* Jan.							+			
Janschinella garetzkii (Merk.)									+	+
Sphenia							1			1
Corbula (*Caryocorbula*)							2			
Lentidium							1		cf	1
Spheniopsis depressa Koen.							+			
S. plana Koen.					+		+			
Gastrochaena langewieschei Gorg.							+			
Panopea woodwardi Koen.										+
Cyrtodaria angusta (Nyst, West.)						+	+		+	+
Panomya koeneni (St.)							+			
Turneria planulata (Gorg.)							+			
Jwannetia neuvillei Coss.							+			
Lyonsia obovata Koen.					+		+			
L. heckeri Merk.										+
Thracia							2			
Poromya hanleyana Wiech.							+			
Cuspidaria subcuspidata Orb.							+			
C. praecuspidata Gil., Th.						+	+			
Cardiomya kochii (Phil.)						+	+			
Verticordia aff. *punctata* Heer.							+			
Number of species	112	60	88	62	62	71	173	33	28	51
Number of genera and subgenera	64	39	73	49	47	56	111	30	25	46

cellian and Egerian time, the zoogeographic affinity of this fauna significantly changed: according to Baldi (1973, 1986), the Kiscellian and early Egerian assemblages were dominated by the north European species (approximately 40 species in the Hungarian assemblages) and "Paratethyan endemics" (about 100 species, gastropods included). Sixty percent of the latter species are known only from Hungary, while others were reported from Central Europe. Species in common with those of the eastern Paratethys are rare.

In the second half of the Egerian, the share of Mediterranean immigrants gradually increased, and in the early Miocene assemblages, they became dominant. The late Oligocene and early Miocene assemblages are not readily discriminated from one another.

The inheritance degree of the late Oligocene fauna of Central Europe from older faunas is low. According to Baldi's estimates, approximately 75 bivalve and gastropod species (30%) are known from the Rupelian time and earlier, 60 species appeared at the beginning of the Chattian, and even more species (about 120) appeared in the earliest Egerian. The number of species that passed into the Miocene is considerably larger (90 species, or 30%) than that of species inherited from the Rupelian.

The composition of the molluscan assemblage of southern Bavaria is also similar to that of the above-considered assemblages (K_p = 0.61–0.72), but is closer to the composition of faunas from the North Sea basin (Tables 1 and 2): typical Mediterranean species were absent here, and dominant species were represented by those of wide distribution, which were characteristic of the northern assemblages. The share of subtropical genera was low (14%), whereas that of endemic species made up 18% (13 species). Other species were represented by forms common to the Chattian assemblages of northern Germany and Belgium.

West Europe. In Aquitaine, Chattian is represented by the Peyrere Beds (data by Czepreghy-Meznerics, 1964; cited after Baldi, 1973). Bivalve mollusks are probably absent or rare in these deposits, and therefore, their composition is unknown. Gastropods are mainly represented by endemic forms, whose degree of inheritance from the Stampian assemblages is very low, whereas the share of species passing into the Miocene is high (one third of the total assemblage, according to O. Amitrov's estimates).

Northward, the Chattian marine sediments were recorded in Belgium, the Netherlands, and Germany. I will start the analysis of these assemblages with the richest and best-studied assemblage from the north German depression, the Chattian stratotypical area. Bivalve mollusks comprise here more than 170 species of 111 genera and subgenera (Table 2, after Görges, 1957; Hinsch, 1972; Janssen, 1979). The lower Chattian assemblages were revised by the latter author. The high diversity of the assemblage and presence of many subtropical genera and species (*Arca biangula, Acar, Atrina, Pinctada, Plicatula, Isognomon, Pteria, Spondylus, Ctena, Chama, Cardita, Crassatina, Arcopagia*, and *Venus*) indicate that it cannot be considered as a boreal one, though it is commonly accepted in the literature. The share of subtropical forms in the assemblage from the Kiscellian sands of the lower Chattian (Janssen, 1979) made up 35%, while other, probably somewhat deeper water assemblages from Germany, Denmark (Harder, 1963), and Belgium (Glibert, 1957) were distinctly dominated by cold-water boreal genera and species, which were widespread in the Oligocene: *Nucula comta, Chlamys bifida, Astarte* (several species), *N. excomatulum, C. beirychi*, and others. When analyzing the inheritance degree of the lower Chattian assemblage (350 bivalve and gastropod species), Janssen noted that 157 species passed from the Rupelian, whereas most bivalve molluscan species (252 species or 56%) first appeared in the Chattian. Only 78 species (17%) of the early Chattian mollusks crossed the Oligocene–Miocene boundary, and bivalves are more diverse in this group than gastropods.

The assemblages from Belgium, the Netherlands (Albrecht and Valk, 1943), and Denmark are noticeably poorer than those from northern Germany, yet with a great number (58) of species in common, they are undoubtedly close to the latter, and the differences between them are probably caused by facies peculiarities.

East Europe. The assemblages from the Askania Formation of the Black Sea region and Kalmyk Formation of the Volga-Don region are impoverished, each containing about 20 species. They almost lack thermophilic subtropical genera and species, while those in common with taxa from the North Sea basin (20 from a total 28 confidently determined species) prevailed. Other species (*Plagiocardium abundans, Cerastoderma prigorovskii, C. karaschokense,* and others) typical of the assemblages from the southern part of the former Soviet Union were endemics of the Eastern Paratethys.

Transcaspian region. Bivalve mollusks of the Karagiin Formation of Mangyshlak are even poorer (16 species), and almost identical to those of the above-considered assemblage. Considerably richer are the assemblages from the Karatomak and Baigubek formations of northern Ustyurt (more than 50 species). They also almost lack species of subtropical genera and are dominated by species common to the assemblages of the northwest European seas. The endemic species represent one third of the Karatomak fauna and a half of the younger Baigubek fauna.

Considerably poorer are the Chattian molluscan assemblages from the Kyzyl Kum and Kopet Dagh areas, but they are mostly represented by the same species with *Chlamys bifida, Plagiocardium abundans*, and *Cerastoderma prigorovskii* as most characteristic. It is interesting to note the appearance of *Megaxinus (Gibbolucina)* and *M. (Saxolucina)* species in the Kopet Dagh molluscan assemblage, both unknown from other assemblages of the Eastern Paratethys.

COMPARISON OF THE ASSEMBLAGES AND ANALYSIS OF GENERA AND SPECIES DISTRIBUTION AREAS

The main difference in the distribution of the late Oligocene molluscan fauna in the west European seas as compared with that in the late Eocene was a considerably higher affinity degree of all the considered assemblages (Table 1). This is a result of smooth climatic barriers and a high degree of panmixia, which resulted in the formation of very extended distribution areas of most genera and a great abundance of species. The high values of Preston's discrimination coefficient (more than 0.86) were obtained only when comparing the northernmost and easternmost faunas with the Tethyan assemblages. The basins of Central Europe were populated by assemblages close in composition to both the Tethyan (northern Italy) and North European faunas. The highest difference is between the molluscan assemblage of Central Europe and that of the Eastern Paratethys (K_p = 0.80–0.87) and its southern Transcaucasian part (K_p = 0.84–0.85), which was close in climatic characteristics to the Tethys. Some of the similarities of these faunas are only at the expense of widespread species (Group IV in Table 2); endemic species are absent from them.

Despite their slight disparities, the North European assemblages show great similarity between each other and those from the southern Bavaria basin (K_p = 0.49–0.56). A large number of species in common, which is reflected in Preston's coefficient values, is also characteristic of the molluscan assemblages of the Norwegian-Greenland basin and Eastern Paratethys (K_p = 0.68–0.80). The similarity between them was particularly high at the beginning of the Chattian, whereas in its second half, the freshening of the Eastern Paratethys caused the extinction of many marine immigrants and an increase in the share of endemics.

The number of species whose distribution areas covered the entire considered region was not less than 24 (Group IV in Table 2; for comparison, only eight such species are known from the Rupelian). Their number could be greater, if the fauna of the Mediterranean region were better studied. However, a large group of the most thermophilic species and ten genera were distributed only in the Mediterranean (Fig. 2) and, on the contrary, many forms are unknown from this region (Groups I, III, V, and VI in Table 2).

Although the assemblages of the central European basins show, as was noted earlier, the transitional features between the Mediterranean and North European faunas, the considerable share in the molluscan forms belonged here to endemic species unknown from the Chattian deposits of other regions (more than 50 species of bivalves and gastropods). Particularly characteristic were also the distribution areas of species that occur in the basins of northwestern and Central Europe and are unknown in the eastern Paratethys. Most widespread among them were *Nucula schmidti, Yoldia glaberrima, Palliolum decussatum*, and *Cuspidaria clava*. Baldi (1973) cited 25 species, which first appeared in the Chattian and were common to both the Central European and North Sea basins.

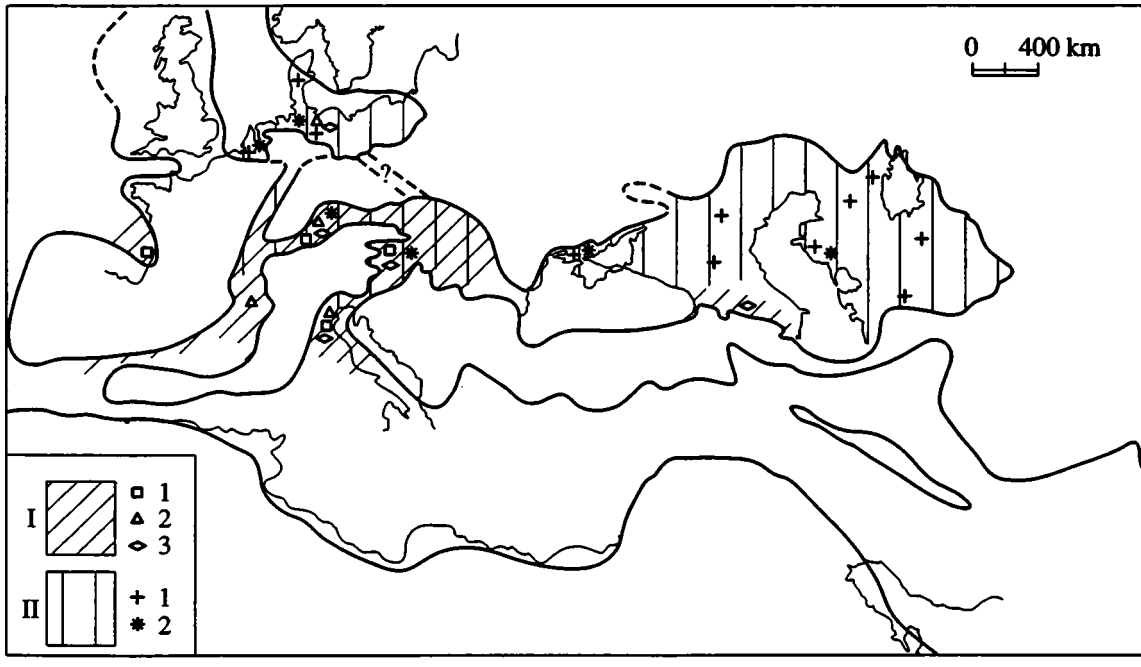

Fig. 2. Distribution areas of some characteristic Chattian molluscan genera and species.
I—Tethyan genera: 1, *Crassatella* s.s., 2, *Spondylus*, 3, *Arca*; II—North European genera and species: 1, *Chlamys bifida*, 2, *Scalaricardita*. Symbols show their collecting sites.

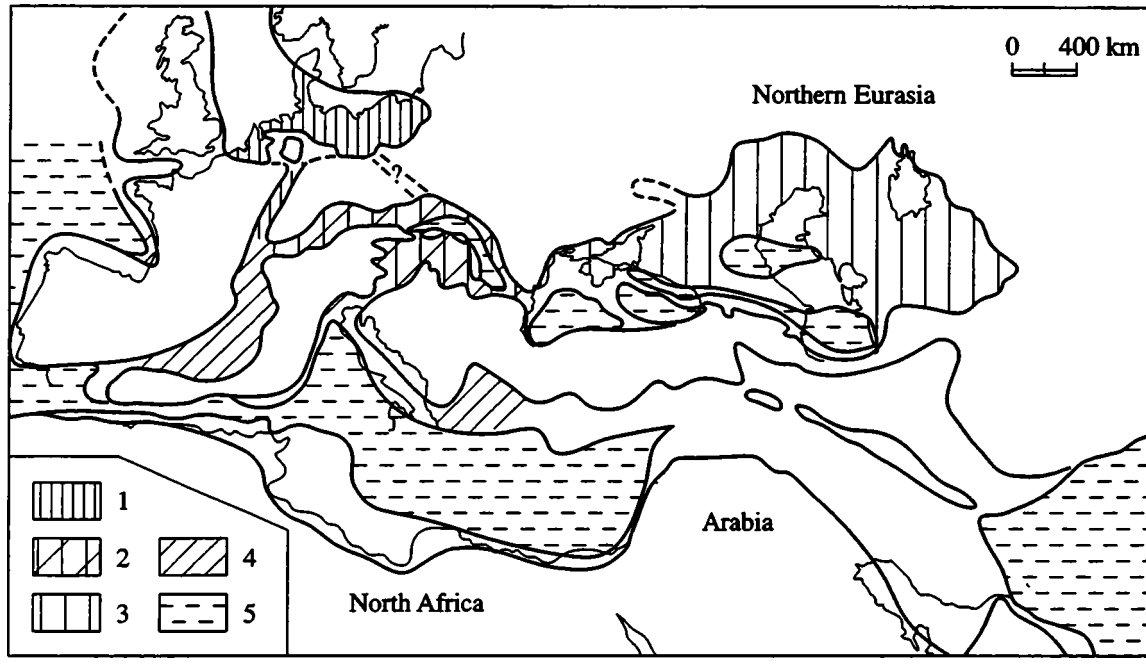

Fig. 3. Zoogeographic division of the West European sea shelves based on the late Oligocene bivalve records.
1, German Province; 2, Central Paratethyan Province; 3, Eastern Paratethyan Province; 4, Tethyan Subprovince; 5, deep-water areas of basins outside the shelf.

There were also many species and some genera whose distribution was limited by the coldest waters of the Norwegian-Greenland Sea, and northern and eastern parts of the eastern Paratethys (Group VI in Table 2). Most characteristic among them is *Clamys bifida*, the Chattian A zonal species, which was abundant and widespread, with distribution area extending to the Kyzyl Kum and Kopet Dagh regions (Fig. 2). A few characteristic dominant species were endemics of the eastern Paratethys (*Plaqiocardium abundans, Cerastoderma prigorovskii, Lentidum kuzhasaicum*, and others).

ZOOGEOGRAPHIC REGIONS

The above-considered data show that during the Chattian, the differences between the Tethyan and North European subprovinces became significantly less pronounced. A wide transitional zone, which appeared between them, cannot be confidently referred to any of these subprovinces. However, the differences in composition of molluscan assemblages of the southern shelf of the Norwegian-Greenland basin and Eastern Paratethys, on the one hand, and the Mediterranean fauna, on the other hand, were still significant (K_p = 0.87–0.90), and these faunas continued to be related to different subprovinces (Fig. 3).

Although an intense faunal exchange between the central European, Tethyan and Norwegian-Greenland basins undoubtedly existed, their biological connections with the Eastern Paratethys were problematic. In this case, it remains unclear as to which way fauna of the northwest European basin used to migrate in the eastern Paratethys at the earliest Chattian time. Among the 30 species in common for the northern part of the eastern Paratethys and northwestern Europe, five widely distributed species (*Nuculana gracilis, Yoldiella pygmea, Chlamys bifida, Ch. hoeninghausi*, and *Palliolum simile*) were absent from the central European basins, evidently, because of their warmer environments. Therefore, it should be assumed that either the faunal migratory routes were northward from the inner Carpathian basins, probably, extending along the outer arc of the Carpathian, or the faunal exchange occurred only at the very beginning of the Chattian, when the fauna of the western Paratethys was most impoverished.

Of all the considered basins, the fauna of the east Paratethyan province was of the most cryophilic, boreal type. In addition to the impoverished composition (compared with the northwest European fauna), it was also featured by the presence of characteristic endemic species. The Transcaucasian part of the eastern Paratethys was, probably, outside this province because its molluscan fauna distinctly differed in both the bivalve molluscan assemblages and their climatic affinity, i.e., in the presence of subtropical genera.

The fauna of the northwest European basin (southern Bavaria basin included) was undoubtedly a part of the single German province, which was characterized by subtropical-boreal molluscan composition.

The Chattian shallow-water faunas of the Central European basins were more heterogeneous in mollus-

can composition, which is evidently an indication of their transitional character and taxonomic changes with time. Only conventionally can they be united into the central Paratethyan province (Baldi, 1986), which, according to the general faunal composition, cannot be referred to either the Mediterranean or North European subprovinces. According to Baldi's data, it probably was a part of the North European subprovince in the first half of the Chattian, but during the Egerian, the influence of the Mediterranean fauna became prevalent here.

ACKNOWLEDGMENTS

The work was financially supported by the Russian Foundation for Basic Research (project 94-04-11187).

Reviewer M.A. Akhmet'ev

REFERENCES

Accordi, B., Stratigrafia e paleontologia delle formationi oligomioceniche del Trvigiane orientale, *Mem. Inst. Geol. Min. Univ. Padova*, 1955, vol. 19, pp. 3–64.

Albrecht, I. and Valk, W., Oligocäne Invertebraten von Süd-Limburg, *Med. Geol. Stichting Maastricht., Ser. G*, 1943, vol. 4, no. 13.

Baldi, T., *Mollusc Fauna of the Hungarian Upper Oligocene* (Egerian), Budapest: Acad. Kiado, 1973.

Baldi, T., *Mid-Tertiary Stratigraphy and Paleogeographic Evolution of Hungary*, Budapest: Acad. Kiado, 1986.

Glibert, M., Pélécipodes et gastropodes du Rupélian supérior et du Chattien de la Belgique, *Mem. Inst. R. Sci. Nat. Belg.*, 1957, no. 137.

Görges, I., Die Mollusken der oberoligozänen Schichten des Deberges bie Bünde in Westfalen, *Paläont. Z.*, 1957, vol. 31, pp. 116–134.

Harder, P., De oligocaene Lag, *Dan. Geol. Unders.*, 1913, vol. 2, no. 22.

Hinsch, W., Mollusken–Biostratigraphie des Moizäns and Chatts im Sädteil des Kreises Herzogtum Lauenburg Schleswig-Holstein, *Meyniana*, 1972, vol. 22, pp. 63–70.

Hölzl, O., Die Mollusken-Fauna der oberbayerischen marinen Oligozänmolasse zwischen Isar und Inn und ihre stratigraphische Auswertung, *Geol. Bavarica*, 1962, no. 50.

Janssen, R., Revision der Bivalvia des Oberoligocäns, *Geol. Abh. Hessen*, 1979, vol. 78.

Popov, S.V., Zoogeography of of Late Eocene Basins of West Europe According to Bivalve Mollusks, *Stratigr. Geol. Korrelyatsiya*, 1994, vol. 2, no. 6, pp. 103–118.

Popov, S.V., Voronina, A.A., and Goncharova, I.A., Stratigraphy and Bivalve Mollusks of the Oligocene–Lower Miocene of the Eastern Paratethys, *Tr. Paleontol. Inst.*, 1993, vol. 256.

Venzo, S., La fauna cattiana delle glauconie bellunesi, *Mem. Inst. Geol. Univ. Padova*, 1937, vol. 13.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 392–402. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 83–94.
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Problems of Eocene Siliceous Phytoplankton Zonation (Exemplified by the Caspian Eocene Deposits)

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Received June 16, 1994

Abstract—Taxonomic studies on diatom and silicoflagellate assemblages from Eocene deposits of the northeastern Caspian region, along with the analysis of stratigraphic and geographic species distribution, confirmed the validity of the North Eurasian provincial zonation elaborated for the subdivision and correlation of marine siliceous deposits. New materials provide grounds to define the taxonomic composition of biozonal assemblages more exactly, establish zonal boundaries at the appearance levels of index-species in epicontinental basins, distinguish some infrazonal subdivisions, and accurately correlate the zones with the standard stratigraphic scale. Despite the occurrence of index-species of some oceanic diatom zones, the insufficient study of diatom phylogenesis prevented us from using them to detect the regional zonation. The refined provincial silicoflagellate zonation based on phylogenetic records was correlated with the more generalized oceanic scale based on these microfossils and standard nannoplankton zonation. Co-occurrences of siliceous plankton and nannoplankton in the Caspian Eocene sections supported the data and substatiated the correlation of established zones with the standard stratigraphic scale.

Key words: *Diatoms, silicoflagellates, Eocene, oceanic and provincial zonal schemes, the Caspian basin.*

The refined Paleogene stratigraphy of the Caspian basin is of both scientific and practical significance because of the occurrence of some important mineral deposits there. The Paleogene deposits are spread in supradome grabens, compensated basins, and interdome depressions. The deposits comprise a variety of facies and contain different microfossil groups, among which diatoms and silicoflagellates are of great importance. The study of their assemblages shows that because of the alternation of carbonate and siliceous deposits, the Eocene sections of the basin, especially in the eastern areas, yield a unique material for correlation between zonal schemes based on siliceous phytoplankton and those based on the Paleogene orthostratigraphic groups of microfossils.

In addition, index-species of some oceanic zones appear successively in the algal assemblages. A significant advantage of oceanic over provincial zonal schemes is that they can be directly correlated with the standard nannoplankton zonation (Fenner, 1985; Bukry, 1981; Perch-Nielsen, 1976, 1985; and others). Provincial zonation based on lit-par-lit studies of the Paleogene sections, which show a facies-dependent absence of planktonic foraminifer and nannoplankton assemblages, can be correlated with the stardard stratigraphic scale only arbitrarily (Gleser, 1979a, 1979b; Bugrova *et al.*, 1991). Therefore, the goals of my investigations were not only to summarize the biostratigraphic studies on the Caspian algae, but also to improve the Eocene provincial zonation, and to find out whether the oceanic zonation can be applied to the regional stratigraphy of platform areas.

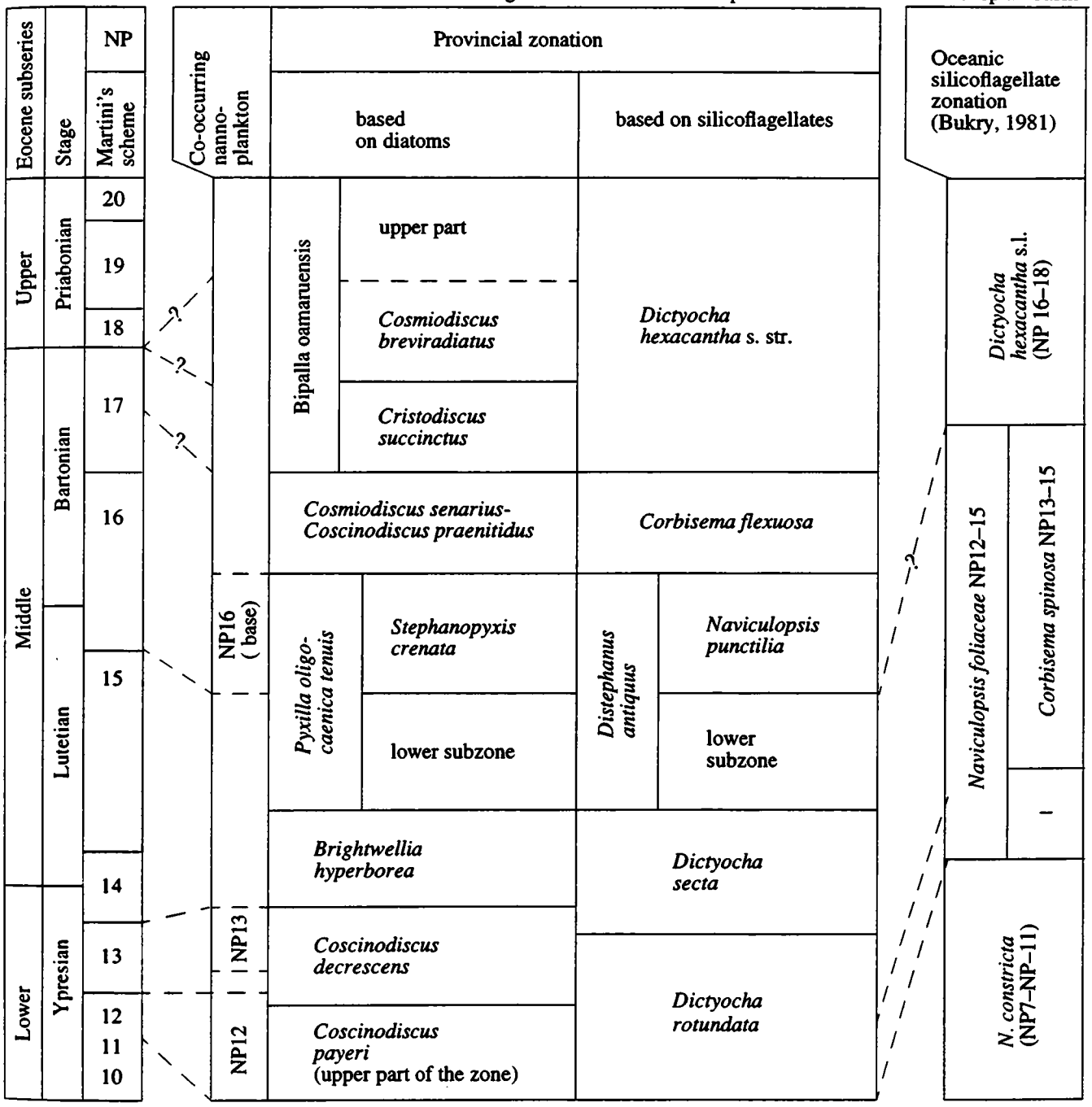
Studies on the Paleogene diatoms and silicoflagellates of the Caspian basin were initiated by my investigations nearly 25 years ago (Gleser, 1967, 1974, 1979a, 1979b; Gleser and Tabachnikova, 1985). Some fragmentary data was published by Ben'yamovskii *et al.* (1990).

This work is based on the revised diatom and silicoflagellate specimens from the collections of S.N. Razmyslova and Yu.P. Nikitina, as well as on the data on these algal groups from the borehole sections sampled by V.N. Beniamovskii and N.S. Tkachenko (Fig. 1). The specimens from the Borehole SP-1 section, kindly submitted by G.V. Larionova, were also studied to more accurately define the composition of siliceous phytoplankton assemblages. The stratigraphic scheme for the Eocene of the Caspian basin (Beniamovskii *et al.*, 1990) was used in the work. Nannoplankton data was communicated by I.P. Tabachnikova.

This article presents the results of lit-par-lit investigations of the Borehole SP-1 section, which exhibits the most complete diatom and silicoflagellate zonal succession. The distribution of these fossils was also studied in other borehole sections. Siliceous phytoplankton assemblages almost entirely characterize the Eocene section of the northeastern Caspian region (Figs. 2 and 3). Six zones and five subzones, based on diatoms, and six zones, based on silicoflagellates, were established there (table).

The studied diatoms and silicoflagellates of the Bailisai Formation were sampled from its stratotype section (Borehole SP-1, depth interval 214–157 m). No algal remains were found in the lower part of the forma-

Zonal correlation schemes based on diatoms and silicoflagellates for the Eocene deposits of the northeastern Caspian basin



tion (interval 214–198 m). Upsection (interval 198–187 m), there are valves of single diatom species and poorly preserved silicoflagellate skeletons.

The diatom *Coscinodiscus payeri* Zone (top) and the silicoflagellate *Dictyocha rotundata* Zone (base) were identified in the upper part of the formation (interval 184–158.2 m). The zonal diatom assemblage consists of *Coscinodiscus decrescenoides* Jouse, *C. payeri* Grun. var. *payeri*, var. *continuus* Streln. et var. *subrepleta* Grun., *Pseudopodosira hyalina* (Jouse) Sheahuk., *P. westii* (W. Sm.) Sheshuk. et Gles., and *Pseudotriceratium chenevieri* (Meist.) Gles. Throughout the zone, there occur *Distephanosira architecturalis* (Brun) Gles., *Riedelia borealis* Sheahuk., and *Pyxilla gracilis* Temp. et Forti. From the base upward the Bailisai Formation section, diatoms increase in diversity mainly at the expense of transitional species known from the upper Paleocene: *Pseudopodosira bella* Possn. et Gles., *Janishia antiqua* Grun., *Trochosiropsis mirabilis* (Kitt.) Gles., *Gruniwiella gemmata* (Grun.) V.H., *Hemiaulus polymorphus* var. *morsianus* Grun., and

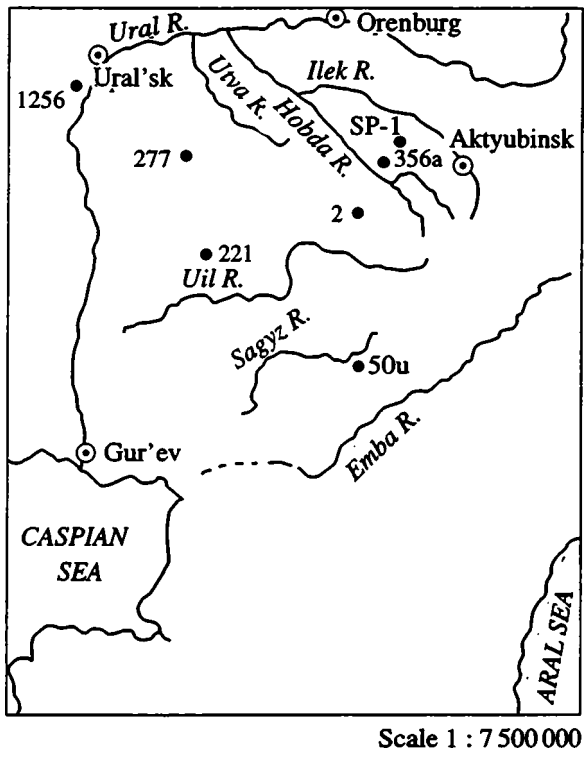


Fig. 1. Location of the studied borehole sections. Points with numbers designate the boreholes.

others. This diatom assemblage is distinguished from the coeval assemblage from the West Siberian lowland, because the latter shows the presence of *Distephanosira architecturalis* and *Riedelia borealis*, as well as a local abundance of *Pyxilla gracilis*, which are not typical of this stratigraphic level.

Until now, the abundance of the latter species was considered to be a characteristic feature of the overlying zones (Gleser, 1979a; Strel'nikova, 1992). For this reason, Strel'nikova (Beniamovskii *et al.*, 1990) established the *Pyxilla gracilis* Zone in this interval. However, this inference did not agree with the diatom and silicoflagellate distribution in the Bailisai Formation. The quantity of valves of *Pyxilla gracilis* appeared to be dependent on enclosing rock properties. Thus, gray deposits contain single or rare valves of this species, and this is also characteristic of the *Coscinodiscus payeri*. The most abundant valves are confined to green clay interbeds: the valves occur in abundance at the depth of 179 m and frequently at the depth of 164 m. It is remarkable that upsection, they demonstrate a decrease instead of an increase in abundance, which is typical of a zone. This distribution of the *Pyxilla pracilis* valves is surely facies-controlled. As noted above, the composition of diatoms suggests them to be the markers of the *Coscinodiscus payeri* Zone, the more so, as they do not include species typical of the *Pyxilla gracilis* Zone.

The silicoflagellate *Dictyocha rotundata* Zone, correlative with the upper part of the *Coscinodiscus payeri* Zone (Bugrova *et al.*, 1991), was established in the same interval (184–158.2 m). The zone yielded peculiar species of the genus *Dictyocha*, with a rectangular basal ring and well-developed radial horns (*D. deflandrei* Freng. var. *deflandrei*, var. *completa* Gles. and *D. frenguellii* Defl. var. *frenguellii*, var. *carentis* Gles.), which associate with *Naviculopsis constricta* (Schulz) Freng. *sensu* Bukry and species with reduced radial horns and a rounded basal ring (*Dictyocha transitoria* Defl. and *D. rotundata* Jouse). *Naviculopsis foliaceae*, which is an index-species of the lower–middle Eocene oceanic zone correlative with nannoplankton zones NP 12–NP 15 (Bukry, 1981), was also found there. This species is distributed up to the Shubarsai Formation. In the silicoflagellate zonation of Martini (1974), *Dictyocha transitoria* is an index-species of the lower Eocene zone, which was distinguished in the stratotype Unter-Eocene-4 section of northern Germany and correlated with Zone NP 13. Later (Martini and Muller, 1976), it was correlated with the top of Zone NP 12 and base of Zone NP 13. Because the Caspian assemblage does not include *Naviculopsis minor*, which is an index-species of the next zone in the Unter-Eocene-4 section, the lower part of the *Dictyocha rotunda* Zone can be regarded as correlative to the upper part of Zone NP 12. The compatibility of the siliceous phytoplankton and nannoplankton zones in this interval of the Borehole SP-1 section supports this correlation.

The *Coscinodiscus payeri* Zone of diatoms was established in full scope in the Borehole 2 section (interval 215–240 m). The upper part of this zone (interval 215–217 m) coincides with the *Dictyocha rotundata* Zone.

The Sholaksai Formation, which is distinguished higher in the Borehole SP-1 section (interval 86–157 m), is characterized entirely, except for a minor interval, by siliceous phytoplankton assemblages, which are indicative of three diatom zones and four silicoflagellate zones.

The *Coscinodiscus decrescens* diatom zone was established in the lower part of the formation (interval 144–156 m) by the change in species composition, with its lower boundary being marked by the appearance of the index-species. The zonal assemblage differs greatly from the previous one: a number of typical species disappear, *Pyxilla gracilis* occurs in abundance throughout the zone, and some new species appear in addition to the index-species. The diatom assemblage is poor at the base of the zone, but is fully represented in its upper part (interval 144–148 m), where species unknown from the lower stratigraphic level are found: *Pyxidicula cruciata* Ehr., *Coscinodiscus mirabilis* Jouse, *C. asteroides* Truan et Witt (?), "*Triceratium*" *basilica* Pant., and *Hemiaulus gurowii* (*grunowii*) Pant. The last species, which was identified by Strel'nikova (1992) as *H. polymorphus* var. *harkovianus* Jouse, is the index-

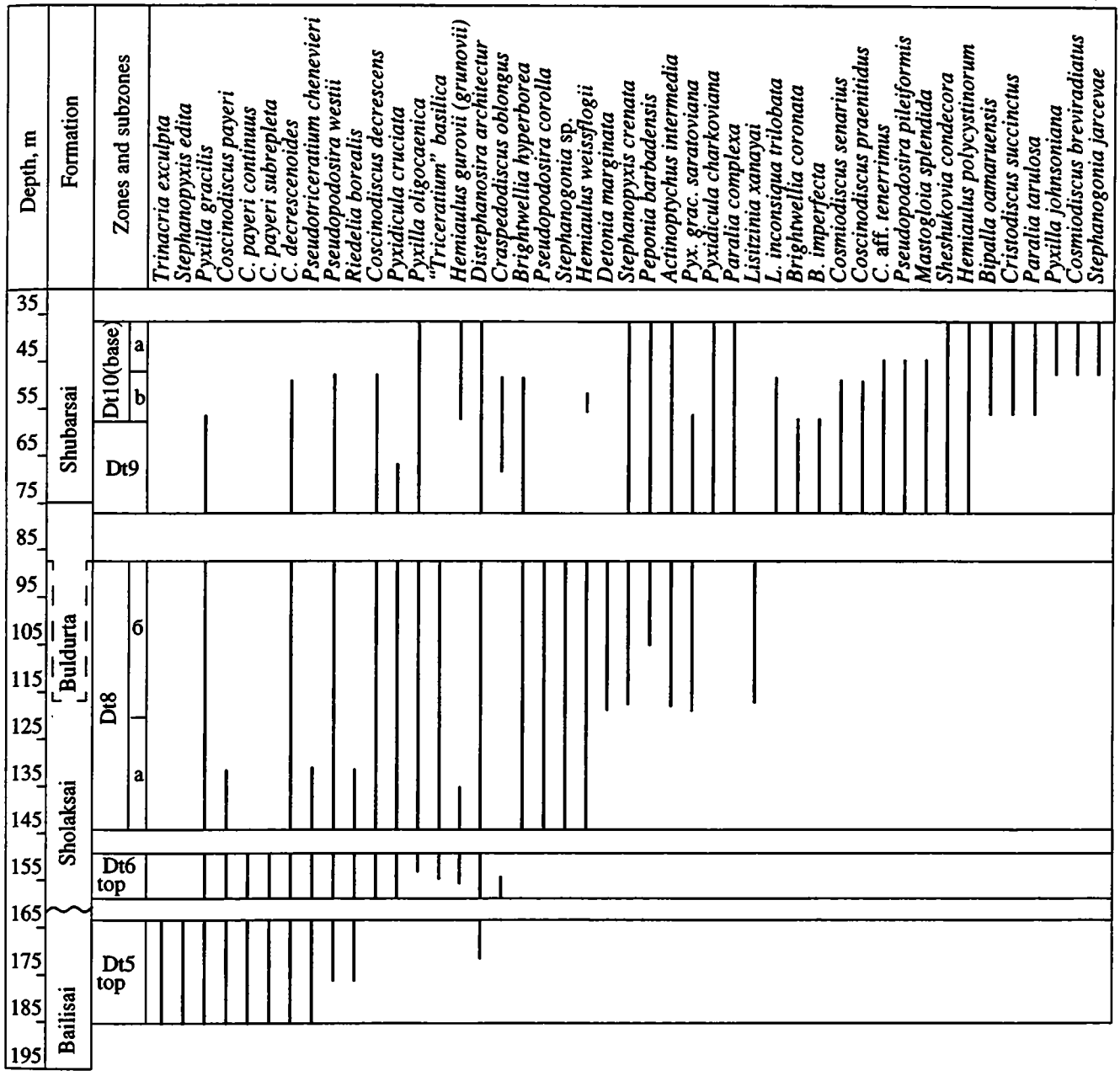


Fig. 2. Zonal diatom assemblages in the Borehole SP-1 section.

species of the beds she distinguished in the middle Eocene Tishki Formation of the Voronezh syneclise. This species appears at a lower stratigraphic level in the Middle Volga section (City of Kuznetsk), from where it was described, and in the Caspian sections. As will be shown below, it appears again in the middle–upper Eocene Shubarsai Formation.

This zonal assemblage, containing single *Pyxilla oligocaenica* var. *tenuis* Jouse and *Craspedodiscus oblongus* Grev., is distinct from the coeval assemblage of the northern regions.

The last form is the index-species of the oceanic zone recognized in the Norwegian Sea, the central and southern parts of the Atlantic ocean, and the Tasman Sea (Fenner, 1985; Strel'nikova, 1992). However, this zone cannot be used in the Caspian Eocene zonation because there are some differences in the taxonomic composition of the diatom assemblages and their position in stratigraphic schemes elaborated for different paleobasins. For instance, in the Bahama region (DSDP Hole 390 A), the partial range of the *Craspedodiscus oblongus* distribution that was first established by Fenner (1985) occupies the lowest stratigraphic position correspondent to that of nannoplankton zones NP 13

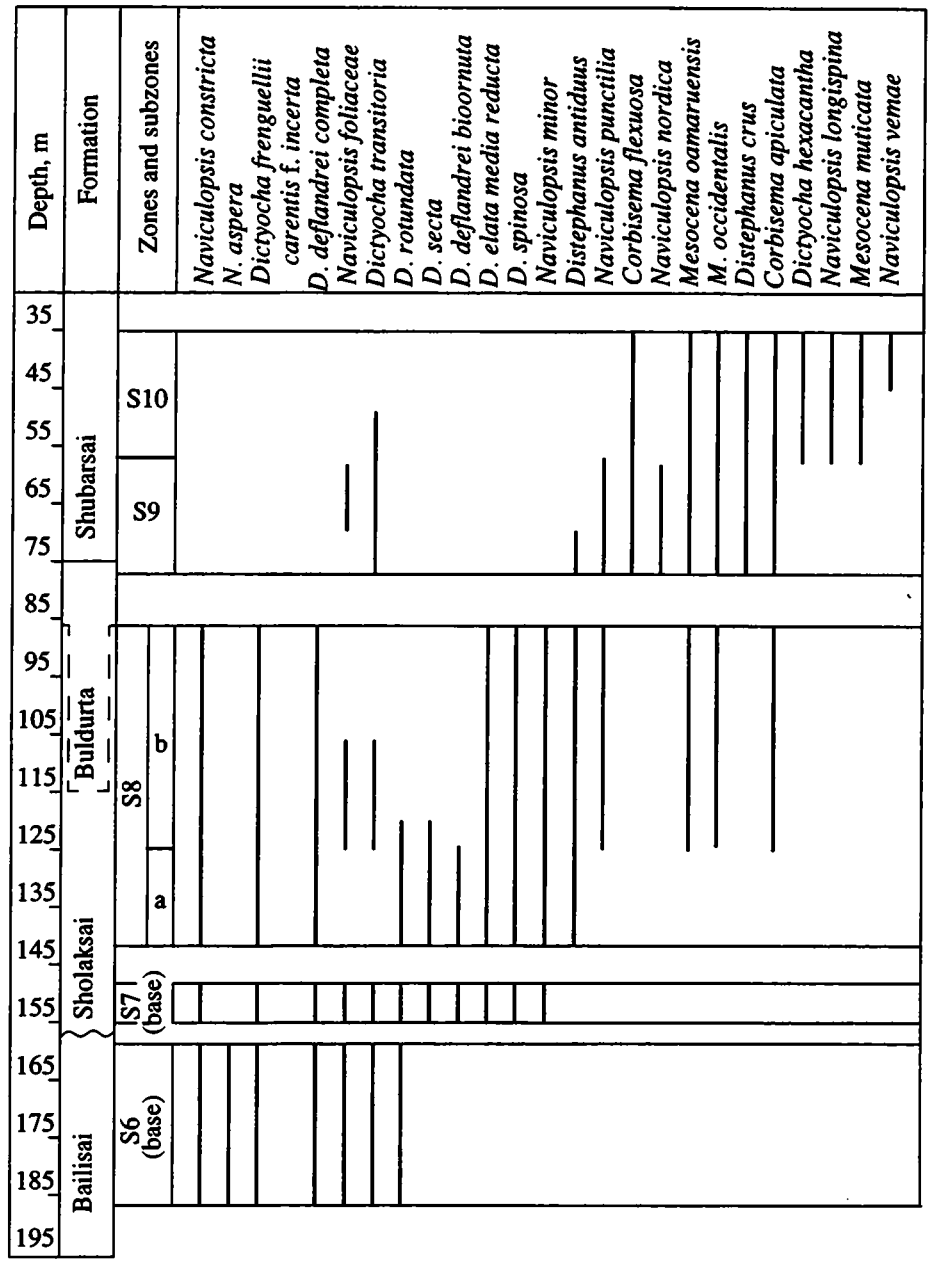


Fig. 3. Zonal silicoflagellate assemblages in the Borehole SP-1 section.

and NP 14 (base). The complex zone of *Craspedodiscus oblongus* in the Norwegian Sea was assigned by this author to the second half of the middle Eocene. According to Strel'nikova's data (1992), this zone is correlative with nannoplankton Zone NP 14 (the lower part of the middle Eocene) in the Norwegian Sea, with the lower part of Zone NP 16 (the second half of the middle Eocene) in the Falkland Plateau, and with the lower part of the Eocene Zone NP 15 of the Tasman Sea. The first appearance level of the index-species, which, according to the interpretation of the cited author, should mark the lower zonal boundary, was not detected in any of the sections, except for DSDP Hole 390 A. Thus, the units, which were defined in different paleobasins, appear to be teilzones of a single biozone of *Craspedodiscus oblongus*, the oceanic species widespread at the end of early and in the middle Eocene. Although this species appears at the same level in the sections of Borehole SP-1 and DSDP Hole 390 A, its sporadic distribution makes it impossible to establish the corresponding partial range zone in the Caspian region.

In the Borehole SP-1 section, the upper part of the diatom *Coscinodiscus decrescens* Zone (interval 144–148 m) corresponds to the silicoflagellate *Dictyocha secta* Zone, whereas its lower part is deprived of silicoflagellates. In the Eocene section of the Kamyskol'

(Borehole 221), the *Coscinodiscus decrescens* Zone (interval 234–310 m) corresponds to the *Dictyocha rotundata* (interval 234–310 m) and *Dictyocha secta* (depth 234 m) zones. These silicoflagellate zones were also distinguished in the sections of boreholes 250u, and 1256.

In addition to the index-species, the assemblage from the upper part of the *D. rotundata* Zone includes *D. elata* var.-*media* f. *reducta* Gles. (this form is known from the higher stratigraphic level in the northern regions), *D. transitoria* Defl., *Corbisema spinosa* Defl., and *Naviculopsis minor* (Schulz) Freng. *Corbisema spinosa* is the index-species of the oceanic zone (or subzone) correlative with nannoplankton zones NP 13–NP 15 (Bukry, 1981). Martini (1974) distinguished the silicoflagellate *Naviculopsis minor* Zone overlying the *Dictyocha transitoria* Zone at the appearance level of this index-species in the Unter-Eocene-4 section, northern Germany. The zone was considered to correspond to Zone NP 13, but it was later correlated to zones NP 13 (top) and NP 12 (base) (Martini and Muller, 1985).

In the Borehole SP-1 section, the diatom *Coscinodiscus decrescens* Zone and the silicoflagellate *Dictyocha secta* Zone were recognized in the interval of nannoplankton Zone NP 13. In the Borehole 221 section, the siliceous phytoplankton zones occur in the lowermost part of the interval bounded by nannoplankton zones NP 13 and NP 16.

Thus, the presented data allows us to correlate the *Coscinodiscus decrescens*, *Dictyocha rotundata* and *D. secta* zones with Zone NP 13 of the terminal lower Eocene.

The next *Brightwellia hyperborea* Zone of diatoms was only recovered in the northern margin of the Caspian basin (Borehole 1256, intervals 269–275 and 209–218 m). The same section (interval 270–280 m) yielded mollusks of the Lutetian age (Barkhatova and Razmyslova, 1974). In the Sholaksai Formation section of Borehole SP-1 (interval 139–144 m), single poorly preserved diatom valves occur above the *Coscinodiscus decrescens* Zone. No diatom remains were found at this stratigraphic level in other sections. The diatom assemblage of the *Brightwellia hyperborea* Zone shows close taxonomic similarity with that of the underlying zone. The boundary between these zones is drawn at the appearance level of *Brightwellia hyperborea* Grun.

It is noteworthy that Gombos (1982) established the *Brightwellia hyperborea* interval corresponding to Zone NP 15 at the appearance level of this species in the Eocene deposits of the Blake–Bahama Plateau (DSDP Hole 390 A). Later, Fenner (1985) included this interval into the upper part of the partial range zone of *Triceratium kanayae*, which corresponded to zones NP 14 (top) and NP 15. Strel'nikova (1992) considered part of this interval to be the *Hemiaulus polymorphus* var. *charkovianus* Zone correlative with Zone NP 15.

Brightwellia hyperborea first appears at the same level as Zone NP 12 in the northeast Atlantic ocean (Fenner, 1988) and in the Blake–Bahama Plateau (Strel'nikova, 1992). In the platform basins, it first appears above the *Coscinodiscus decrescens* and *Dictyocha rotundata* zones, i.e., above Zone NP 13. In the Borehole 1256 section, the *Brightwellia hyperborea* Zone corresponds to the silicoflagellate *Dictyocha secta* Zone (top).

The discussed *Coscinodiscus decrescens* and *Brightwellia hyperborea* diatom zones overlying successively the *Coscinodiscus payeri* Zone correspond approximately to the earlier established local complex zone of *Pyxilla gracilis* (Gleser, 1979a; Bugrova *et al.*, 1991). The easily detectable and widespread Eocene species *Pyxilla gracilis* serves as the index-species of this zone in different zonation schemes. Although uniformly named, these zones are based on different principles and characterized by different zonal assemblages, ranges, stratigraphic positions, and geographic extent.

So, in the Lower Ob' region, Rubina and Droznes (1961) distinguished zones on the basis of the quantity of valves of the index-species ("zone with frequent *Pyxilla gracilis*", and "zone with abundant *Pyxilla gracilis*"). As shown above, the quantity of valves is facies-controlled and, hence, such zones can only be used as local. The above-mentioned local complex zone of *Pyxilla gracilis* is characterized by a specific diatom species composition different from those of overlying and underlying zones, includes abundant *Pyxilla gracilis* throughout, and has interregional significance (Gleser, 1979a; Bugrova *et al.*, 1991). It was assigned to the upper part of the lower and lower part of the middle Eocene, and was approximately correlated with nannoplankton zones NP 13 and NP 14.

A subglobal zone of the middle Eocene was established by Jouse (1979) based on the appearance of two index-species. It should be noted that this zone is incorrectly cited in literature as the *Pyxilla gracilis* Zone (Fenner, 1985; Fourtanier, 1991; Strel'nikova, 1992). According to Fenner's data, this zone represents the terminal early–initial middle Eocene. Strel'nikova distinguished the *Pyxilla gracilis* Zone based on the principle suggested by Jouse—the appearance of the index-species. In her opinion, it was confined to the underlying *Coscinodiscus payeri* Zone. In this case, the scope of the *Coscinodiscus payeri* Zone accepted by Strel'nikova (1992) remains unclear. She considers the *Pyxilla gracilis* Zone to be correlated with the lower Eocene radiolarian zones and with nannoplankton Zone NP 12.

Actually, the first appearance of *Pyxilla gracilis* is recorded within the *Coscinodiscus payeri* Zone in the most Eocene sections of different regions of northern Eurasia (Gleser, 1979a). However, rare valves of this species are also known from the underlying *C. josephinus (uralensis)* Zone, and even from the upper Paleocene deposits (the Trueva Beds) of the Middle Volga region.

Fourtanier (1991) established the *Pyxilla gracilis* Zone by the appearance of the index-species in the lower Eocene sediments of the Indian ocean. Integrated studies revealed that the species appearance coincided there with the lower limit of Chron C 24, which was determined within the interval of 56.3–56.5 Ma. This zone can be directly correlated with nannoplankton zones CP 9a (the upper half), CP 9b, and CP 10 (in part), which were also distinguished in this section [zones NP 10 (top), NP 11, and NP 12 (in part), respectively, in the standard scheme of Martini].

Thus, in the Indian Ocean, the *Pyxilla gracilis* Zone occupies a lower stratigraphic position than the zone of the same index-species in northern Eurasia. Additionally, *Craspedodiscus undulatus*, which marks its upper boundary, and other characteristic species were not found in northern Eurasia. All of this implies that we are dealing with two different zones named after the same index-species.

Taking into account the controversial information on various characteristics of the *Pyxilla gracilis* Zone, it seems expedient to distinguish two zones instead—the *Coscinodiscus decrescens* and *Brightwellia hyperborea* zones. This will make the earlier proposed diatom zonation more refined and precise.

Upsection of the Sholaksai Formation that was recovered by Borehole SP-1 (interval 139–186 m), the diatom *Pyxilla oligocaenica* var. *tenuis* and silicoflagellate *Distephanus antiquus* zones are recognized. There was no essential taxonomic renewal in these zonal assemblages, as compared with those of the underlying zones. The assemblages consist mainly of the transitional species that appeared in the *Coscinodiscus decrescens* and *Brightwellia hyperborea* zones. The diatom assemblage is dominated by *Pyxilla oligocaenica* var. *tenuis*. Other species occur either in abundance [*Coscinodiscus decrescens*, *Brightwellia hyperborea* Grun., *Pseudopodosira hyalina* (Jouse) Sheshuk., and *P. corolla* (A.S.) Streln.] or sporadically [*Hemialus gurovii* (grunowii), and *H. weissflogii* Pant.]. Species *Stephanogonia* sp., and *Actinoptychus intermedia* A.S., which are present throughout the zone, should be mentioned among of the first occurring forms. The older transitional species *Pseudopodosira bella* Poss. et Gles., *Distephanosira architecturalis*, and *Pyxilla gracilis* are still of importance there. Apparently, the lower zonal boundary should be drawn at the disappearance level of *Coscinodiscus payeri* varieties.

The silicoflagellate assemblage of the *Distephanus antiquus* Zone contains the dominating index-species and some transitional species that occur throughout: *Dictyocha rotundata*, *D. secta*, *D. elata* var. *media* f. *reducta*, *Mesocena elliptica*, *Corbisema spinos*, and *Naviculopsis foliaceae*. The upper parts of the *Pyxilla oligocaenica* var. *tenuis* and *Distephanus antiquus* zones (interval 86–125 m) represent the diatom *Stephanopyxis crenata* and silicoflagellate *Naviculopsis punctilia* subzones.

Although the algal assemblages retain many species of the underlying zones, there are changes in their specific and generic compositions due to the appearance of a number of new taxa. Among the diatoms, these are *Pyxidicula charkoviana* (Jouse) Streln. et Nikol., *Paralia complexa* (Lohman) Andrews, *Actinoptychus intermedia* A.S., *A. pericavatus* Brun, *Coscinodiscus subtilis* Ehr., *Peponia barbadensis* Grev., *Stephanopyxis crenata* Sheshuk., *Lisitzinia inconspiqua* var. *trilobata* (Fenner) Gles., *Detonia marginata* Gles., *Pyxilla gracilis* var. *saratoviana* (Pant.) Temp. et Forti, *Sheshukovia sparsimpunctata* (Jouse) Gles., and *Rutilaria areolata* Sheshuk. The listed species, except for the index-species, are of limited distribution and occur sporadically.

It should be noted that for the reasons discussed above, it appears to be impossible to recognize the oceanic *Craspedodiscus oblongus* Zone at this stratigraphic level in the Caspian sections, despite the frequent occurrence of this species.

The silicoflagellate assemblage of the *Naviculopsis punctilia* Zone comprises both transitional species (*Distephanus antiquus*, *Dictyocha elata*, var. *media* f. *reducta*, *Dictyocha spinosa*, and *Naviculopsis foliaceae*), which are still of great importance, and some new forms (*Mesocena oamaruensis* Schulz, *M. oamaruensis* var. *apiculata* Schulz, *Corbisema ovalis* Perch-Nielsen, and *Naviculopsis punctilia* Perch-Nielsen). *Dictyocha rotundata* and *D. secta* disappear at the top of the subzone.

According to the data of Perch-Nielsen (1985), in the oceanic sections, *Mesocena oamaruensis*, *Corbisema ovalis*, and *Naviculopsis punctilia* appear at the lower boundary of the *Dictyocha hexacantha* Zone together with the index-species. Bukry (1981) determined the boundaries of this zone by the appearance and disappearance of the index-species. The zone represents the stratigraphic interval corresponding to nannoplankton zones NP 16–NP 18 (base). In the studied sections of the Caspian region, *Dictyocha hexacantha* appears at a higher level than the listed zonal species. So, if the appearance level of zonal species is taken as a base, the lower boundary of the oceanic zone must coincide with that of the *Naviculopsis punctilia* Subzone in the Borehole SP-1 section. If the appearance of *Dictyocha hexacantha* is taken into account, this subzone should be correlated with the underlying oceanic *C. spinosa* Subzone corresponding to nannoplankton zones NP 13–NP 15. It should be noted that Bukry refers *Dictyocha secta* and *Distephanus antiquus*, which are the index-species of provincial zones, to the assemblage of the *Naviculopsis foliacea*.

The *Pyxilla oligocaenica* var. *tenuis* and *Distephanus antiquus* zones were recognized, in addition to the Borehole SP-1 section, in Boreholes 221 (depth 224 m), 2 (interval 65–110 m), 356a (interval 65–95 m), 50u (interval 167–197 m), and 1256 (interval 164–203 m).

The *Stephanopyxis crenata* and *Naviculopsis punctilia* subzones were established in the upper parts of these zones in the sections of boreholes SP-1, 2 (interval 65–81 m), 50u (interval 167–170 m), 356a (interval 65–75 m), and 1256 (interval 164–167 m).

This analysis of the silicoflagellate stratigraphic distribution is supported by the fact that in the Borehole 1256 section, the algae of the *Stephanopyxis crenata* and *Naviculopsis punctilia* subzones were found together with the nannoplankton of the lower part of Zone NP 16.

Data on the diatom and silicoflagellate zonation of the Sholaksai Formation dates the formation as the terminal early–mid-middle Eocene, and suggest it to be approximately correlative with nannoplankton zones NP 13–NP 16 (base). According to Ben'yamovskii *et al.* (1990), the Sholaksai Formation includes only one diatom *Pyxilla gracilis* Zone of the lower Eocene (as it is interpreted by Strel'nikova). The lower part of the formation was assigned on the basis of radiolarian, benthic foraminifer and nannoplankton records to the upper part of the lower–lower part of the middle Eocene.

The Buldurta Formation recovered by Borehole SP-1 (interval 75–85 m) was not studied in terms of algae distribution, except for the uppermost part (interval 75–77 m). Two diatom zones and two silicoflagellate zones were established in the stratotype section of this formation (Borehole 271, interval 237–270 m; Ben'yamovskii *et al.* 1990). The largest, lower part of the formation (interval 240–267 m) corresponds to the *Stephanopyxis crenata* and *Naviculopsis punctilia* subzones.

The overlying diatom *Coscinodiscus praenitidus–Cosmiodiscus senarius* and silicoflagellate *Corbisema flexuosa* zones represent the upper part of the Buldurta Formation (3 m thick in the stratotype, and 2.5 m thick in the Borehole SP-1 section) and the lower part of the parastratotype of the Shubarsai Formation (Borehole SP-1, interval 55–75 m).

The specific and generic compositions of the diatoms from this zone differ greatly from those of the underlying subzone because of the appearance of many new floral elements. *Radialiplicata costulata* Gles. and *Monobrachia simplex* Schrader appear together with the index-species *Coscinodiscus praenitidus* Fenner and *Cosmiodiscus senarius* (Grev.) Gles. et Olst. at the lower boundary of the respective zone, and *Hemiaulus polycystinorum* Ehr. and *Hyalodiscus frenguellii* Hanna were found somewhat above the boundary. In different localities, the zone sporadically contains *Brightwellia coronata* (Brightw.) Ralfs, *B. imperfecta* Jouse, *Corona retimervis* Sheshuk. et Gles., *Paralia complexa* (Lokman) Andrews *Pseudotriceratium coscinoides* (Gr. et St.) Gles., *Sheshukovia condecora* (Brightw.) Gles., *Lisitzinia kanayae* (Fenner) Gles., *Pseudopodosira pileiformis* Jouse, *Mastogloia rutilans* Brun, *Hemiaulus crenatus* Grev., and *H. reflexiapinosus* Ross. Many of these species are widespread in the middle Eocene deposits of the Atlantic ocean.

It is remarkable that the new floral forms include index-species of some middle Eocene oceanic zones established by different authors: *Lisitzinia* (*Triceratium*) *kanayae* is the index-species of the lower zone of the middle Eocene, and *Brightwellia imperfecta* is the index-species of the upper zone of the middle Eocene in the low latitudes, according to Fenner's zonation (Fenner, 1985). In Jouse's zonation (Jouse, 1979), *Paralia complexa* (*Melosira fausta* A.S.) is the index-species of the initial middle Eocene zone, and *Hemiaulus polycystinorum* is the index-species of the terminal middle–upper Eocene zone. According to Strel'nikova (1992), *Trinacria subcapitata* (Grev.) Grun. is one of the index-species of the lower subzone of the upper Eocene in the Voronezh anteclise.

Throughout the *Cosmiodiscus senarius–C. praenitidus* Zone, there occur *Pyxilla oligocaenica* var. *tenuis* (this species is represented by single or rare specimens in the Buldurta Formation, and dominates again in the Shubarsai Formation), single or rare *Lisitzinia incomapiqua* var. *trilobata*, and single *Craspedodiscus oblongus*.

Despite the presence of index-species of the oceanic zones, it appears to be impossible to recognize the latter in the Caspian sections, as the assemblages and the index-species marking zonal boundaries are different.

The silicoflagellate *Corbisema flexuosa* Zone was also established in this part of the Shubarsai Formation. The assemblage of this zone essentially differs from that of the *Naviculopsis punctilia* Subzone. Along with the index-species, many new species appear, e.g., *Mesocena occudentalis* Hanna et Bukry, which occurs throughout the zone, and *Naviculopsis nordica* Bukry and *Distephanus crux* (Ehr.) Hack, both represented by sporadic single specimens. Some transitional species still exist: *Mesocena oamsruensis*, *Naviculopsis punctilia*, *N. constricta*, *N. foliaceae*, *Dictyocha spinosa*, *Distephanus antiquus* (the last form abundance is reduced to single specimens), and *Corbisema apiculata* (Lemm.) Hanna. The latter species is known from the lower Paleocene. In the oceanic zonation (Perch-Nielsen, 1985), its disappearance determines the upper boundary of the *Corbisema apiculata* Zone of the uppermost Eocene–basal Oligocene beds, which overlies the *Dictyocha hexacantha* Zone.

According to the data of Perch-Nielsen (1985), the above-listed new species, except for *Distephanus crux*, appear in the middle–upper Eocene oceanic sections at the same level as *Dictyocha hexacantha*, i.e., at the lower boundary of the *Dictyocha hexacantha* Zone. The zone corresponds to nannoplankton zones NP 16–NP 18. In the studied Eocene sections of the Caspian region, *Dictyocha hexacantha* appears at a higher level than the mentioned zonal species.

In addition to the Borehole SP-1 section, the *Cosmiodiscus senarius–Coscinodiscus praenitidus* and *Corbisema flexuosa* zones were established in the sections of Boreholes 2 (interval 45–64 m), 271 (interval 199–

240 m), and 356a (interval 35–65 m). The *Corbisema flexuosa* Zone was also identified in the Borehole 50u section (depth 130 m).

The Eocene zonation (Gleser, 1979a; Bugrova *et al.*, 1991) is completed by the *Bipalla* (*Melosira*) *oamaruensis* Zone. New data permits the further subdivision of this zone, comprising two subzones in its lower part.

Upsection of the Shubarsai Formation (Borehole SP-1, interval 54–60 m), the *Cosmiodiscus senarius–Coscinodiscus praenitidus* Zone of diatoms is replaced by the *Cristodiscus succinctus* Subzone of the *Bipalla oamaruensis* Zone. The lower boundary of these zonal units is determined by the appearance of the index-species *Bipalla oamaruensis* and *Cristodiscus succinctus*, as well as by *Paralia torulosa* Gles. and occasional *Cristodiscus duple* (Gles.) Gles. et Olst.

The zone is characterized by new elements, which include some species of the two young genera *Bipalla* and *Cristodiscus*, and by a great amount of transitional species. *Cristodiscus* (*Coscinodiscus*) *succinctis* is an index-species of particular beds, which Strel'nikova (1992) distinguished in the deposits of the second half of the middle Eocene (within the Zone NP 16) in the Blake–Bahama Plateau and in the Tishki Formation of the Voronezh syneclise.

Boundaries of these beds were not established in the first locality, whereas in the second one and in the Caspian region, their lower boundary was drawn at the appearance level of *Cristodiscus* (*Coscinodiscus*) *succinctus* and *Bipalla* (*Paralia*) oamaruensis. Some other species appearing at this boundary in the Voronezh syneclise were found at the lower (*Stephanopyxis* (*Pyxidicula*) *crenata*, *Peponia barbadensis* Grev., and *Corona retinervis* Sheshuk. et Gles.) or higher (*Xanthiopyxis oblonga* Ehr.) stratigraphic levels in the Caspian Eocene section.

In the Caspian region, the lower boundary of the *Cristodiscus succinctus* Subzone of the *Bipalla oamaruensis* Zone coincides with the appearance level of *Dictyocha hexacantha*, the index-species of the middle–upper Eocene oceanic zone. According to Bukry (1981), the oceanic *Dictyocha hexacantha* Zone is equal in scope to the biozone of the index-species, whose appearance and disappearance determine the lower and upper zonal boundaries, respectively. Because some species of this zone appeared in the Caspian region much earlier than the index-species, it may be supposed that the level of appearance of *Dictyocha hexacantha* in oceans is not coincident with that in the region. Most likely, the oceanic zone corresponds to the *Naviculopsis punctilia* Subzone and the *Corbisema flexuosa* and *Dictyocha hexacantha* s.s. zones of the provincial zonation.

The *Criatodiscus succinctus* Subzone and the corresponding part of the *Dictyocha hexacantha* s.s. Zone were also recognized in the Borehole 271 section (interval 179–199 m).

Upsection of the Shubarsai Formation (Borehole SP-1, interval 37–54 m), the *Cosmiodiscus breviradiatus* Subzone of diatoms corresponding to the silicoflagellate *Dictyocha hexacantha* s.s. Zone was distinguished. The lower subzonal boundary was drawn at the appearance level of the index-species.

The specific character of this diatom assemblage is emphasized by the appearance of new forms (*Cosmiodiscus breviradiatus* Gles. et Olsht. *Stephanogonia jarcevae* Sheshuk., *Xanthiopyxis globosa* Ehr., *X. oblonga* Ehr., and sporadic *Pyxilla* cf. *prolongata* Brun *sensu* Jouse).

The assemblage is also characterized by the disappearance of some species that are stratigraphically important for the underlying zones (*Cosmiodiscus senarius*, *Coscinodiscus praenitidus*, *Craspedodiscus oblongus*; and *Pyxilla gracilis*, the latter disappearing at the top of the lower zone). There are many transitional species, which still retain their importance: *Bipalla oamaruensis*, *Paralia complexa*, *P. torulosa*, *Radialiplicata costulata*, *Pyxidicula charkoviana*, *Actinoptychus intermedia*, *Peponia barbadensis*, *Hemiaulus gurovii* (*grunowii*), *H. polycystinorum*, *Distephanosira architecturalis*, *Pyxilla oligocaenica* var. *tenuis*, *P. gracilis* var. *saratoviana*, *Cristodiscus succinctus*, etc. The *Cosmiodiscus breviradiatus* Subzone corresponds to a part of the *Dictyocha hexacantha* s.s. Zone. The silicoflagellate assemblage is almost identical to that discussed above, except for the occurrence of single specimens of *Naviculopsis longispina* and *Mesocena muticata* Gles., and the complete disappearance of *Distephanus antiquus*.

The *Cosmiodiscus breviradiatus* Subzone and the *Dictyocha hexacantha* s.s. Zone were also identified in the sections of boreholes 221 (interval 140–155 m, 9 m above the lower Boundary of nannoplankton Zone NP 17), 356a (interval 10–30 m), and 2 (interval 7–15 m). This suggests that this subzone can be equally correspondent to the upper part of the middle Eocene and to the upper Eocene.

The upper part of the stratotype of the Shubarsai Formation (interval 27–37 m) was not studied with respect to diatom and silicoflagellate distribution. The upper parts of the *Bipalla oamaruensis* and *Dictyocha hexacantha* s.s. zones that were recognized in the Borehole 221 section (interval 70–110 m) contained foraminifers, which Ben'yamovskii assigned to the upper Eocene *Globigerapsis tropicalis* Zone. The algae are poorly preserved, and the assemblages are very impoverished. More representative assemblages with distinctive features were found at the depth of 70 m.

The diatom assemblage does not include new forms, except for *Asterolampra marylandica* Ehr. represented by single specimens. There are predominantly middle Eocene transitional species: *Pyxidicula charkoviana*, *Sheshukovia condecora*, *Pseudopodosira pileiformis*, *Xanthiopyxis oblonga*, *Pyxilla oligocaenis* var. *tenuis*, etc. The disappearance of *Cosmiodiscus breviradiatus*

and *Cristodiscus succinctus*, both occurring throughout the underlying *Cosmiodiscus breviradiatus* Subzone, is noteworthy. *Asterolampra marylandica* is the index-species of the oceanic zone of the uppermost part of the middle Eocene–lower part of upper Eocene, as it was established by Fenner (1985) in the low latitudes. The possibility of correlating a part of the *Bipalla oamaruensis* Zone with the *Asterolampra marylandica* Zone cannot be excluded, but this correlation cannot be made on the basis of the data. At the same time, the disappearance of *Lisitzinia inconspiqua* var. *triloba* marks the lower boundary of the *Asterolampra marylandica* Zone at the low latitudes. Fenner reported on the disappearance of *Crespedodiscus oblongus* and *Brightwellia hyperborea* within this zone.

According to the data of McLean and Barron (1984), *Lisitzinia inconspiqua* var. *trilobata* and *Craspedodiscus oblongus* do not cross the middle–upper Eocene boundary in the California sections. Fenner (1988) also believed that the lack of *Craspedodiscus oblongus* and *Lisitzinia* (*Triceratium*) *inconspiqua* var. *trilobata* in the diatom assemblage of the *Coscinodiscus* aff. *tenerrimus* Zone of the Norwegian Sea to be indicative of its late Eocene age. Strel'nikova (1992) assigned this zone to the first half of the middle Eocene. In the Caspian sections, *Craspedodiscus oblongus* disappears at the lower boundary of the *Cosmiodiscus breviradiatus* Subzone, while *Brightwellia hyperborea* disappears slightly above it.

The silicoflagellate assemblage comprises the species typical of the *Dictyocha hexacantha* Zone, except for the new species *Naviculopsis vemae* Perch-Nielsen. According to Perch-Nielsen (1976, 1985), this form is the index-species of the lower subzone of the *Dictyocha hexacantha* Zone in the bottom sediments of the Norwegian Sea. The upper subzone is defined by the disappearance of *Naviculopsis punctilia*. In the Caspian Eocene sections, *N. punctilia* disappears well before the first occurrence level of *N. vemae*. It is remarkable that Perch-Nielsen (1976) considers the middle–upper Eocene boundary in the Norwegian Sea sections to be coincident with the lower boundary of the *N. vemae* Zone, which occurs somewhat higher than that of the *Dictyocha hexacantha* Zone.

The biostratigraphic distribution of diatoms and silicoflagellates in the Caspian region suggests that the middle–upper Eocene boundary is likely inside the *Bipalla oamaruensis* and *Dictyocha hexacantha* s.s. zones.

The *Bipalla oamaruensis* Zone, confined to the Obukhovka Formation of the Dnieper–Donets basin, was previously dated as late Eocene on the basis of combined paleontological records (Gleser, 1979a). However, the lower part of this zone may belong to the uppermost part of middle Eocene.

The boundary between the two Eocene subseries cannot be defined more accurately until siliceous plankton assemblages are found together with planktonic foraminifers or nannoplankton. In my opinion, the Shubarsai Formation can only be preliminary referred to the terminal middle–upper Eocene. Possibly, the Lutetian–Bartonian boundary lies at the base of this formation in the case when it includes the clayey member, 2.5 or 3 m thick, which is now tentatively referred to the Buldurta Formation composed mainly of sands, silts, and siltstones.

Thus, the detailed studies of the diatom and silicoflagellate assemblages allowed the possibility to refine the Eocene zonal biostratigraphy of siliceous deposits in the northern and eastern Caspian regions. The previous provincial zonations based on these algal groups were improved, and zonal and infrazonal boundaries were established at the first appearance levels of species, which are of stratigraphic importance for the North Eurasian provinces.

So, the *Pyxilla gracilis* Zone, with an uncertain lower boundary, was replaced by the *Coscinodiscus decrescens* and *Brightwellia hyperborea* zones. The *Pyxilla oligocaenica* var. *tenuis* Zone was subdivided into two parts: a lower subzone and the *Stephanopyxis crenata* Subzone. The new *Cosmiodiscus senarius*—*Coscinodiscus praenitidus* Zone was established. The lower half of the *Bipalla oamaruensis* Zone was subdivided into the *Cristodiscus succinctus* and *Cosmiodiscus breviradiatus* subzones.

In the silicoflagellate-based zonation, the upper part of the *Distephanus antiquus* Zone was identified as the *Naviculopsis punctilia* Subzone, and the new *Corbisema flexuosa* Zone was established. The correlation of the provincial oceanic zonation schemes allowed a more precise age determination of these zones.

In most cases, the appearance levels of oceanic zonal index-species in the Caspian sections cannot serve as datum-planes. This is because the diatom zonations, including oceanic ones (Fenner, 1985), are of provincial rank, and their first appearance levels of species mark the moment of favorable environments in a definite water area. The distribution of the studied Eocene diatom and silicoflagellate assemblages in the marine epicontinental basins of northern Eurasia implies that the Caspian basin belonged to the Euro-Siberian algal paleoprovince at the terminal early–initial middle Eocene, and to the Euro-Turgai province in the second half of middle–late Eocene.

Because provincial ecological environments are controlled by a complex of global and local factors determing the algal flora character, a more reliable correlation should be based on the taxonomic similarity of zonal assemblages rather than on datum-planes.

REFERENCES

Barkhatova, N.N. and Razmyslova, S.N., *Stratigrafiya i nummulitidy eotsenovykh otlozhenii Prikaspiiskoi vpadiny* (The Eocene Stratigraphy and Nummulitids of the Caspian Basin), Leningrad: Nauka, 1974.

Beniamovskii, V.N., Segedin, R.A., Akopov, T.R., *et al.*, New Paleocene and Eocene Formations of the Caspian Basin, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1990, vol. 65, issue 5, pp. 68–75.

Bugrova, E.M., Tabachnikova, I.P., Nikolaeva, I.A., *et al.*, The Paleogene System, in *Zonal'naya stratigrafiya fanerozoya SSSR* (The Phanerozoic Zonal Stratigraphy of the USSR), Moscow: Nedra, 1991, pp. 130–142.

Bukry, D., Synthesis of Silicoflagellate Stratigraphy for Maestrichtian to Quaternary Marine Sediments, *Spec. Publ. Soc. Econ. Paleontol. Mineral.*, 1981, vol. 32, pp. 433–444.

Fenner, J., Late Cretaceous to Oligocene Planktic Diatoms, *Plankton Stratigraphy*, Bolli, H.M., Saunders, J.B., and Perch-Nielsen, K., Eds., Cambridge: Cambridge Univ., 1985, pp. 713–762.

Fenner, J., Occurrences of Pre-Quaternary Diatoms in Scandinavia Reconsidered, *Meyniana*, 1988, no. 40, pp. 133–141.

Fourtanier, E., Paleocene and Eocene Diatom Biostratigraphy and Taxonomy of Eastern Indian Ocean Site 752, *Proc. Ocean Drill. Program: Sci. Results*, Washington: Ocean Drilling Program, 1991, vol. 121, pp. 171–187.

Gleser, Z.I., Occurrence of Late Eocene Diatoms, Silicoflagellates, and Ebridians at the Western Mugodzhary Slope, in *Novosti sistematiki nizshikh rastenii* (News on Lower Plant Systematics), Leningrad: Nauka, 1967, pp. 21–25.

Gleser, Z.I., The Eocene Diatoms, in *Diatomovye vodorosli SSSR: Iskopaemye i sovremennye* (Diatom Algae of the USSR: Fossil and Recent), Leningrad: Nauka, 1974, vol. 1, pp. 115–143.

Gleser, Z.I., Paleogene Diatom Zonation, *Sov. Geol.*, 1979a, no. 11, pp. 19–30.

Gleser, Z.I., Paleogene Silicoflagellate Zonation for the Mediterranean Area within the Limits of the USSR, *Plankton i organicheskii mir pelagiali v istorii Zemli, Trudy XIX sess. Vsesoyuz. paleontol. o-va* (Plankton and Organic World of the Pelagic Zone in the History of the Earth, Proc. XIX Sess. All-Union Paleontol. Soc.), Leningrad: Nauka, 1979b, pp. 24–42.

Gleser, Z.I. and Tabachnikova, I.P., Paleogene Biostratigraphy of the North and East Caspian Basin, *Sov. Geol.*, 1985, no. 8, pp. 56–64.

Gombos, A.M., Early Middle Eocene Diatom Evolutionary Events, *Bacillaria*, Braunschweig: Cramer, 1982, pp. 225–242.

Jouse, A.P., Diatom Biostratigraphic Zones of the Eocene, *Nova Hedwigia*, 1979, no. 64, pp. 427–440.

Martini, E., Silicoflagellate Zones in the Eocene and Early Oligocene, *Senckenbergiana Lethaea*, 1974, vol. 54, pp. 527–532.

Martini, E. and Muller, C., Eocene to Pleistocene Silicoflagellates from the Norwegian-Greenland Sea (DSDP, Leg 38), *Initial Rep., Deep Sea Drill. Project*, Washington: US Gov. Print. Off., 1976, vol. 38, pp. 857–896.

McLean, H. and Barron, J.A.E., Late Middle Eocene Diatomite in Northwestern Baja California Sur Mexico: Implication for Tectonic Translation, in *Paleogene Stratigraphy, West Coast of North America*, Los Angeles: Filewicz and Squires, 1988, pp. 1–7.

Perch-Nielsen, K., New Silicoflagellates and a Silicoflagellate Zonation in North European Paleogene and Eocene Diatomites, *Bull. Geol. Soc. Da.*, 1976, vol. 25, pp. 27–40.

Perch-Nielsen, K., Silicoflagellates, *Plankton Stratigraphy*, Bolli, H.M., Saunders, J.B., and Perch-Nielsen, K., Eds., Cambridge: Cambridge Univ., 1985, pp. 813–846.

Rubina, N.V. and Droznes, M.A., The Eocene Deposits of the Lower Ob' Region, in *Resheniya i trudy mezhvedomstvennogo sovechchaniya po dorabotke i utochneniyu stratigraficheskikh skhem Zapadno-Sibirskoi nizmennosti* (Resol. and Proc. Interdepartm. Conf. on Improvement of Stratigr. Schemes of the West Siberian Lowland), Leningrad: Gostoptekhizdat, 1961, pp. 264–271.

Strel'nikova, N.I., *Paleogenovye diatomovye vodorosli* (The Paleogene Diatom Algae), St. Petersburg: St. Petersburg Univ., 1992.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 403–408. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 95–100.
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Correlation of Marine and Terrestrial Pliocene–Pleistocene Deposits in the Western Turkmenistan

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Received June 2, 1994; in final form, December 25, 1994

Abstract—Correlation problems concerning marine and terrestrial deposits of Turkmenistan are considered using new and previously published data on Pliocene and Pleistocene fauna. The studied remains represent different genera, species and subspecies of six Mammalia orders. Changes in the late Pliocene–Pleistocene fauna composition are presumably related to paleogeographic reorganizations of that time in western Turkmenistan. The correlation scheme comprising various faunal assemblages is suggested as reasonably valid for the entire Caspian region.

Key words: *correlation, fauna, Mammalia, Mollusca, Pliocene, Pleistocene, Turkmenistan, Azerbaijan.*

Western Turkmenistan is a very favorable region for elucidating correlative relations between marine and terrestrial deposits, and consequently, very important for refining the stratigraphic scale applicable throughout the Caspian region as a whole. For many years, we studied both types of Pliocene and Pleistocene sections in this region. The validity of our stratigraphic scheme is based on detailed field observations, and substantiated by comprehensive analysis of rich assemblages of brackish-water mollusks and fossil vertebrate fauna. The study was most explicit in southwestern Turkmenistan. This area, extending from the Kara Kum Lowland to the Caspian shoreline, comprises abundant outcrops of the alluvial-lacustrine sequence, which consists of pale and pinkish silts and clays representing deposits of the ancient Amu Darya River. The typical units of the sequence are the beds (0.5–2 m thick) of gray micaceous sands enclosing pale-pinkish silt and clay balls derived from earlier deposits of the Amu Darya River at the stage when its flow rate was accelerated. The sequence bearing the freshwater molluscan fauna (*Unio* and other forms) is known as the Kara Kum Formation. Its deposits are penetrated by drill holes in the eastern part of the Kara Kum Lowland.

In western (several localities) and eastern (one locality) Turkmenistan, the upper horizon of micaceous sands yielded fossil remains of vertebrates. The western localities are confined to the ancient Amu Darya valley. Among these, the most significant is the locality of the Khudai-Dag depression, where Fedorov (1945) discovered a burial of the fossil elephant. Later, the cranium and some other skeletal remains from this burial were described as the holotype of the new species *Palaeoloxodon turkmenicus* Dubrovo (Dubrovo, 1955, 1960). During the geological survey of the 1950s, Fedorov (1959) and Nigarov (in 1962) discovered bone fragments of fossil elephants and artiodactyls in deflation hollows of the Uzboi valley and Urundzhik depression, respectively. Because of the fragmentary character of the remains, they were not identified very precisely, only as *Elephas* s. l. and *Bos* s. l.

In order to obtain more complete data on the Pleistocene vertebrates of this region, the Geological Survey Department of the Turkmenistan and Paleontological Institute (Academy of Science of the USSR) collaborated on a relevant project in the 1980s. Their field expeditions studied the region between the Bol'shoi Balkhan in the north and Kyuren-Dag in the south, and between the Yaskha Lake in the east and Caspian Sea in the west. Our collections at that time included remains representing, though sometimes fragmentary, six orders of mammals, two of birds, one of reptiles, and two of fishes. All of the studied vertebrate localities occur in the alluvial-lacustrine deposits.

The middle Pleistocene fossils are the most abundant. They have all been distinguished in the horizon of micaceous sands bearing clay balls.

In the Uzboi Valley, southwestward of the Yaskha Lake, we found small fragments of elephant tusk and dental plates.[1] Judging from the 3-mm-thick enamel layer, the remains represent either *Achidiskodon*, or *Palaeoloxodon* genera. Here, we also distinguished the eighth marginal tooth plate typical of the *Testudinoidea* species, and a small metapodium fragment of a miniature Artiodactyla form. These remains were collected from light-gray fine-grained micaceous sands, bearing clay pebbles and the freshwater mollusks *Unio* sp., *Corbicula flumimalis* Mull., and *Limnaea* sp. of the

[1] See Dubrovo (1960) for the method to investigate the dentition of fossil elephants.

middle Pleistocene age. In addition to the *in situ* burial of *Palaeoloxodon turkmenicus* in the Khudai-Dag depression, we found nearby an upper molar of this species, dental fragments of *Elasmotherium*, a piece of a deer horn, and vertebrae of sheatfishes *Silurus* sp. From the depression slope, we collected the dental fragment of an elephant representing the genus *Mammuthus*. The coexistence of the genera *Palaeoloxodon* and *Mammuthus* in a given territory is a well-known fact, although these taxa used to reside in different biotopes: the former in forest areas, and the latter in open landscapes. The co-occurrence of both elephant genera in Turkmenistan is in good agreement with the character of deposits bearing their remains. The lithologic features of the sands enclosing the clay balls point to the intense activity of past river flows, and consequently, to gallery forests in their valleys.

Remains of the Turkmeni elephant were also reported from the Urundzhik locality, where bones of *Elasmotherium* were recovered in the anatomic arrangement suggesting the *in situ* type of burial. Artiodactyls are represented here by the Bovidae and Cervidae families. Additional finds include carapace plates of terrapins (Testudinoidea) and cranium fragments of the sheatfish *Silurus* sp. In Cheleken Island, the same horizon yielded dental fragments of the so-called *khozarian* elephant *M. trogontherii chosaricus* Dubrovo. Perissodactyls from the island are represented by three genera: *Elasmotherium*, *Dicerorhinus*, and *Hemionus*; artiodactyls here include two forms of miniature Bovidae, the species *Bos* s. l., and two taxa of Cervidae. In this locality, we also distinguished the bones of hares and carnivores, plates of Testudinoidea, fish vertebrae (*Silurus* and *Esox*), and bird bones of taxa *Falcolinus*, *Anax*, and *Anser* (Dubrovo and Nigarov, 1990). The most significant find was that of *M. trogontherii chosaricus*, which is the basic form of the Khazarian[2] fauna. Its holotype was reported from the middle Pleistocene deposits of the Chernyi Yar locality in the Volga region. Remains of the more recent *trogontherii* subspecies are known in many other middle Pleistocene burials.

The stratigraphic position of the Kara Kum Formation is well defined. In the Kara Kum lowland, it is overlain by marine deposits bearing *Didacna ebersini* Fed. and *D. cristata* Bog. of the early Khvalinian age. In the Cheleken locality, the horizon with clay balls underlies yellow–gray sands enclosing the Khvalinian assemblage of *D. praetrigonoides* Nal. et Anis and *D. pallasi* Prav. (Nevesskaya, 1956). In the Balkhan region and along the Uzboi River, the upper part of the Kara Kum Formation contains large *Unio* shells and remains of *M. trogontherii chosaricus* and some other mammalian forms, implying the middle Pleistocene age of this interval. In Cheleken Island and near Krasnovodsk, pale and pinkish clays in the lower part of the formation yielded the marine mollusks *Didacna subpyramidata* Prav., *D. lindleyi* (Dasch.) Fed., and *D. paleotrigonoides* Fed. (Nevesskaya, 1958; Fedorov, 1953, 1959). Somewhat further to the west, the alluvial-lacustrine sequence is replaced in its total range by marine deposits bearing the Khazarian assemblage of bivalves.

Below the Kara Kum Formation are the Urundzhik Beds, representing the topmost unit of the Bakunian sequence. The unit (1–2 m thick) consists of gray sands bearing abundant mollusks *Didacna eulachia* Bog. (Fed.), *D. pravoslavlevi* Fed., *Dreissena rostriformes* Desh., and *Dr. polymorpha* (Pall.) (Nevesskaya, 1956, 1958; Fedorov, 1953). This horizon is traceable in western Cheleken Island and in the Urundzhik and Kumdag depressions. It has also been recovered by the drill hole near the Yaskha Lake. In the Cheleken, Urundzhik, and Kumdag localities, as well as on the Monzhukly Height and in some dill holes of eastern Turkmenistan, the lower Bakunian strata bear shells of *Didacna parvula* Nal. and *D. catillus* Eichw. typical of this stratigraphic interval. The lowermost unit of the Bakunian sequence consists of coarse-grained cross-bedded alluvial sands of the Tyurkyan Formation. In the Balkhan region, the contact between these sands and underlying Apcheronian deposits corresponds to the erosion surface marked by an angular unconformity.

The Bakunian deposits of Cheleken Island, which underlie the precisely dated lower Khazarian beds with molluscan fauna, yielded indeterminable bone fragments of mammals. It is likely that dental fragments of the horse *E.* cf. *sussenbornensis* Wust and two teeth of a large vole (either the evolutionary late species of the *Mimomys* genus, or the early *Arvicola* form according to determination of A.K. Agadzhanyan) were derived from these deposits, although we found them in a talus. In addition, Pavlova (1926) reported on a horn fragment of *Cervus elaphus* L. discovered in the upper Bakunian strata.

Remains of the Apcheronian terrestrial fauna also occur near the Monzhukly Height. The Apcheronian deposits are represented here by alternating members of clay, silt, and sand. Shingle beds and lenses at their base yielded egg shell fragments of the ostrich *Struthio asiaticus* M. Edw. (Mikhailov and Kurochkin, 1988), and dental plates of an elephant. A thick (3–3.5 mm) enamel and frequency index of the plates (less than 4) suggest that the remains can be attributed to the southern form *Archidiskodon meridionalis* (Nesti). It is impossible to identify the form at the subspecies rank, because the material is too fragmentary. The noted taxon includes two subspecies characteristic of different faunal assemblages (Dubrovo, 1991). The Apcheronian–Tamanian assemblage includes the later subspecies of the southern elephant, i.e., *A. meridionalis*

[2] As in the case of *Eguus chosaricus* Gromova, the name is the Latin transliteration of the relevant Russian word given in the *Glossary of Russian Language* (Ozhegov, 1953). In contrast, the term Khazarian, signifying the geological age or faunal assemblage, is derived from the Khazarian Sea, meaning the Caspian Sea in Turkish.

tamanensis Dubrovo, whose holotype was reported from deposits of respective age in the Sea of Asov region. The described find is the first one in Turkmenistan, and may suggest that *A. meridionalis tamanensis* also existed here during the Apcheronian time.

In Cheleken Island, marine deposits of the Apcheronian time yielded remains of the dolphin species *Delphinus delphis* L. and *Phocaena* sp. (Kirpichnikov, 1951; Ryabinin, 1908).

Older Akchagylian deposits are widespread in western Turkmenistan. They bear marine Mediterranean-type fauna, e.g., *Cardium dombra* Andr., *Mactra subcaspia* Andr., and other forms. The upper Akchagylian beds also enclose *Dreissena polymorpha* Pall., indicating a phase of sea water freshening during the regression.

The Akchagylian vertebrate fauna of southwestern Turkmenistan is detectable by its footprints (Amanniyazov *et al.*, 1979). The middle Akchagylian sandstones of the Kyuren-Dag region (the Ufra, Gyaurli, and Ak-Oba localities in the southern Krasnovodsk Peninsula) display well-preserved footprints of camels, small artiodactyls, carnivores, and birds, who visited the seashore at that time. In addition, Fokanov (1976) described the great gerbil *Pliorhombomys gromovi* Fokanov, whose teeth were found in the Gokcha Formation of the Akchagylian age. Near the Kopet Dagh, the middle–upper Pliocene deposits yielded the remains of Canidae, *Pseudomeriones* sp., and *Testudo horsfieldi* Cray (Birman *et al.*, 1971). Egg shell fragments of *Struthio asiaticus* were detected in the upper Pliocene deposits of the Bodkhyz Bay and Krasnovodsk Peninsula regions (Mikhailov and Kurochkin, 1988).

In eastern Turkmenistan, fossil remains of elephants were first discovered in 1989 during the construction of the Zeid water storage basin. Here, in the northeastern wall of the excavation, Nigarov described the following section as distinctly divisible into three units:

(1) The upper member of light-gray fine-grained eolian sands, whose thickness is 1.7 m.

(2) The Obruchev Formation (in terms of Turkmeni geologists), composed of alternating fine- to very fine-grained and highly calcareous (16–50%) clays. Their distribution area and inner structure suggest them deposited by the North Afghan rivers; the thickness is 9.5 m.

(3) The lower member of fine-grained micaceous sands bearing small brown clay balls and shells of *Unio* sp. and *Anodonta* sp. These sands enclose elephant remains and are 2 m thick in the exposed section.

During the excavation, the cranium and mandible were destroyed. Preserved remains include left and right penultimate upper molars (M^2), three plates of the last upper molar (M^3), and two partially broken plates of presumably the last lower molar (M_3), all investigated by Dubrovo.[3]

Description. The total plate number in M^2 was 13 or 14 (the Plate, Fig. I). The dental crown is narrow and high, with the plate frequency index equal to 6. The enamel layer is 3 mm thick, the sinus is poorly developed, and the dental cementum is unsatisfactorily preserved.

Judging from the plate forms, the crown of M^3 was narrow and high. The three preserved plates are 78, 78, and 80 mm thick and 151, 140, and 171 mm high, respectively; they display the distinct grinding pattern of ·—· type.

The crown of M_3 was likely narrow, with 3-mm-thick enamel. One preserved plate provides the possibility to detect grinding patterns of the tooth. Its apical tubercles are separated by a shallow sinus, while the sinus between the external and internal lateral columellae is deep (the Plate, Fig. II). Further grinding would expose on the mastication surface two lateral disks and a long intermediate oval (·—·), i.e., the grinding pattern typical of the genus *Palaeoloxodon*.

The narrow and high dental crown, along with the characteristic grinding patterns and median, though poorly pronounced, sinus suggest that the studied teeth can be attributed to the species of the *Palaeoloxodon* genus.

Comparing remains from the Zeid locality with other *Palaeoloxodon* forms (Dubrovo, 1981), we cannot refer them to either *P. antiquus* (Falc. et Cautl.), or *P. namadicus* (Falc. et Cautl.), as these forms had somewhat wider dental crowns. Moreover, in M^3 of *P. meridionaloides* (Gromova), dental plates are less frequent, and the enamel is thinner. Thinner enamel layers are also typical of *P. germanicus* (Stefănescu) and *P. naumanni* (Makiyama). The holotype of *P. turkmenicus* is represented only by the posterior molars of a very large mature male. As it is known, these teeth of elephants are larger than M^2, and have more frequent plates and a thicker enamel layer. We noted precisely this difference between M_3 of the *P. turkmenicus* holotype and M^2 of a younger and smaller, presumably female form, which was found in 1989. The enamel layer of M_3 is 3 mm thick in both cases, and the degree of sinus development is also similar. All of these data indicate that the described teeth belong to the fossil elephant *P. turkmenicus*. Their description supplements the known diagnostic features of the genus, and the find refines previous ideas on its distribution area.

The co-occurrence of upper and lower molars of the elephant species suggests that they have not been redeposited. The *Palaeoloxodon turkmenicus* skeleton, which was excavated in the Balkhan region of Turkmenistan, was also buried at the place where the animal died (Fedorov, 1945; Dubrovo, 1955). Both burials are

[3] The materials are in the depository of the Chardzhou Regional Museum of Turkmenistan.

ACKNOWLEDGMENTS

This study was supported by the Russian Foundation for Basic Research (project no. 94-04-13030) and International Science Foundation (grant JAB-100).

Reviewer L.A. Nevesskaya

REFERENCES

Amanniyazov, K.A., Nigarov, A.N., and Uzakov, O.U., Fossil Footprints of Animals in the Ancient Desert, *Probl. Osvoeniya Pustyn* (Ashkhabad), 1979, vol. 5, pp. 80–82.

Birman, A.S., Zhegallo, V.I., Rastsvetaev, L.F., *et al.*, Occurrence of Pliocene Vertebrates in Molassic Deposits of Eastern Kopet Dagh, *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1971, vol. 46, issue 2, pp. 99–106.

Dubrovo, I.A., New Data on the Anatomy and Distribution of Ancient Elephants (*Hesperoloxodon*), *Dokl. Akad. Nauk SSSR*, 1955, vol. 10, no. 4, pp. 759–762.

Dubrovo, I.A., Ancient Elephants of the USSR, *Tr. Paleontol. Inst. Akad. Nauk SSSR*, 1960, vol. 85, pp. 1–78.

Dubrovo, I.A., Die Fossilen Elephanten Japans, *Quartärpaläontol.*, 1981, vol. 4, pp. 49–84.

Dubrovo, I.A. and Nigarov, A.N., Plio–Pleistocene Fossil Vertebrate Localities of Southwestern Turkmenia, USSR, *Quartärpaläontol.*, 1990, vol. 8, pp. 35–45.

Dubrovo, I.A., Stratigraphic Significance of Fossil Elephantinae and the Boundary between the Neogene and Quaternary Systems, *Izv. Akad. Nauk SSSR, Ser. Geol.*, 1991, no. 10, pp. 9–18.

Fedorov, P.V., Occurrence of Vertebrate Skeleton in the Uzboi Lower Courses, *Tr. Turkm. Geol. Upravl.*, 1945, no. 1, pp. 84–85.

Fedorov, P.V., Quaternary Mollusks of the Genus *Didacna* Eichwald from the Caspian Region and their Stratigraphic Significance, in *Stratigrafiya chetvertichnykh otlozhenii i noveishaya tektonika Prikaspiiskoi nizmennosti* (Stratigraphy of Quaternary Deposits and Neotectonics of the Caspian Lowland), Moscow: Akad. Nauk SSSR, 1953, pp. 112–130.

Fedorov, P.V., Stratigraphy of Quaternary Deposits and History of the Caspian Sea Evolution, *Tr. Geol. Inst. Akad. Nauk SSSR*, 1957, vol. 10, pp. 1–296.

Fedorov, P.V., Quaternary Deposits of Western Turkmenistan and Their Position in the Unified Stratigraphic Scale of the Caspian Region, *Tr. Geol. Inst. Akad. Nauk SSSR*, 1959, vol. II, pp. 5–45.

Filippova, N.Yu., Upper Pliocene–Middle Pleistocene Palynological Assemblages from Eastern Transcaucasia and Western Turkmenistan: Implications for Stratigraphy and Paleogeography, *Cand. Sci. (Geol.–Mineral.) Dissertation*, Moscow: Geol. Inst. Ross. Akad. Nauk, 1993, pp. 1–26.

Fokanov, V.A., New Genus and Species of *Pliorhombomys gromovi* Fokanov sp. nov. (Mammalia, Rodentia) from Upper Pliocene Deposits of Southern Turkmenistan, in *Evolyutsiya gryzunov i istoriya formirovaniya ikh sovremennoi fauny* (Evolution of Rodentia and Formation History of Their Recent Fauna), Leningrad: Nauka, 1976, pp. 122–124.

Kirpichnikov, A.O., Dolphins from Apcheronian Deposits, *Dokl. Akad. Nauk SSSR*, 1951, vol. 79, no. 6, pp. 1021–1024.

Lebedeva, N.A., Geological Position of Terrestrial Vertebrate Remains of the Khaprovian, Tamanian, and Tiraspolian Assemblages in the Marine Akchagylian and Apcheronian Beds of Eastern Transcaucasia, *Byull. Kom. Izuch. Chetvertichn. Perioda*, 1972, no. 38, pp. 99–115.

Mikhailov, K.E. and Kurochkin, E.N., Egg Shell of Fossil *Struthioformes* from the Paleoarctic Regions and Their Significance for the Hypothesis of Ratitae Evolution, *Tr. Sovmestnaya Sov.-Mong. Paleontol. Eksped.*, 1988, issue 34, pp. 43–65.

Nevesskaya, L.A., Biostratigraphy of Marine Quaternary Deposits in Turkmenistan *Byull. Mosk. O-va. Ispyt. Prir., Otd. Geol.*, 1956, vol. 31, no. 3, pp. 65–79.

Nevesskaya, L.A., Quaternary Marine Mollusks of Turkmenistan, *Tr. Paleontol. Inst. Akad. Nauk SSSR*, 1958, vol. 65.

Pavlova, M.V., *Cervus tschelekensis* n. sp. and *Alces maeoticus* n. sp., *Ezheg. Russ. Paleontol. O-va*, 1926, vol. IV, pp. 45–72.

Ryabinin, A.N., Dolphin Remains from Pliocene Deposits of the Cheleken Island, *Izv. Geol. Kom.*, 1908, vol. 27, no. 148, pp. 517–521.

Samsonov, S.K., *Paleogeografiya Zapadnoi Turkmenii v novokaspiiskoe vremya* (Paleogeography of Western Turkmenistan during the Novocaspian Time), Moscow: Akad. Nauk SSSR, 1963.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 409–415. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 101–108.
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BRIEF COMMUNICATIONS

Stratigraphy of Mesozoic-Cenozoic Deposits in the Salt Dome Basin of the Peri-Caspian Depression

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Received April 1995

Abstract—Unique natural structures, the secondary (compensated) basins, with thick Mesozoic-Cenozoic suprasalt deposits were considered in the salt dome Pericaspian depression. The place of anomalous basins in the halokinetic development of the depression was defined, and specific aspects of their identification, types, and zoning were described. A lithologic-facial analysis and comparative analysis of rock thickness and sedimentation rate in secondary basins and standard sections were performed. It was concluded that anomalous basins are of particular significance for solving stratigraphic problems of the suprasalt sequence of the Pericaspian depression.

Key words: *Pericaspian depression, salt dome basin, compensation basin, halogenesis, Mesozoic-Cenozoic.*

The Pericaspian depression is a vast salt dome basin, in which halokinetic processes were most intense. These processes highly complicated the structure of the salt-bearing sequence and overlying suprasalt deposits, thus creating difficulties for stratigraphic research. What is the essence of the respective problems?

Salt tectonics is known to affect the size and distribution of uplift and subsidence zones in the sedimentation basin by forming local provenances and controlling the sedimentation rate (sedimentary thickness).

Within the salt dome basin, each dome with contiguous basins presents a certain autonomous sedimentary cell, or halokinetically balanced area, which comprises: (1) a halokinetic uplifted zone (salt accumulating zone), (2) a subsidence zone (salt emigration zone), and (3) a neutral zone, in which the halokinetic component is missing. During the halokinetic process, the dimensions of the zones and their location change. At the early stage (that of "salt pillow"), the halokinetically balanced area actually consists of two zones: (1) a zone of salt accumulation (dome), and (2) an emigration zone (compensated basin), a neutral line (strip) separates them. At the second ("diapir") stage (beginning after the salt masses, capable of migration, were displaced from the primary compensated basin into the dome zone), the structure of the halokinetically balanced area changes. The neutral strip abruptly widens at the expense of the former compensated basin, and forms the so-called interdome zone. The zone of the former dome arch splits into two parts: (1) the salt stock, with continuing salt accumulation, and (2) the secondary compensation basin, where the salt outflow into adjacent salt stock is compensated by subsidence.

Thus, depending on the stage of halokinesis, two (dome, basin), or three (dome, interdome zone, and basin) section types simultaneously develop in each halokinetically balanced elementary zone. *Dome sections* have a reduced thickness and numerous hiatuses. Some of these hiatuses are traceable in adjacent interdome zones, though most of them are local, marking individual features of the dome growth. *Basin sections* are commonly complete and show an anomalously great thickness. No appreciable hiatuses in sedimentation are observed. *Interdome sections* are generated within the interdome zone, from which masses, capable of migration, were squeezed out earlier; with respect to completeness and thickness, these sections occupy an intermediate position. Unconformities and hiatuses recorded in them correspond to the stages of the area regional rising.

The mentioned structural features of sections in different structural zones of the halokinetically balanced area, which become still more complicated as different halokinesis stages occur in each elementary area at different times, present difficulties for interpretation at the very beginning of the research. The purpose in this case is to define standards (stratotypes) of certain stratigraphic subdivisions of the local scale.

According to *Stratifraficheskii kodeks SSSR* (1977), a stratotype section should comply with the following requirements: (a) it should be typical of a given stratigraphic subdivision, (b) it should be complete, without substantial stratigraphic hiatuses, but with a clear sequence, making it possible to observe boundaries with the underlying and overlying deposits (p. 46). Basin sections best comply with these requirements. Owing to the permanent and active subsidence of compensated basins and sufficient amount of sediments to

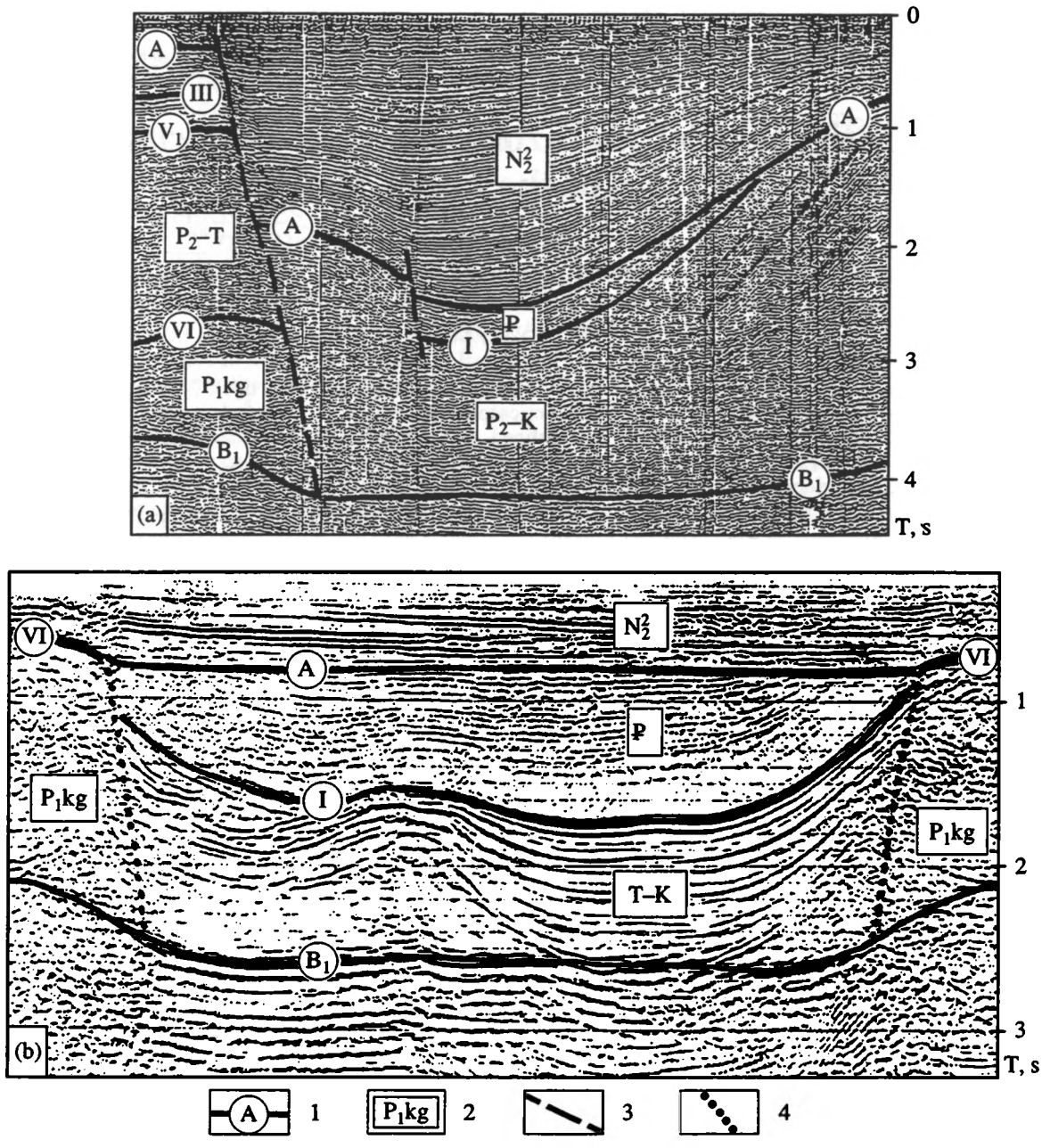

Fig. 1. Seismic section of anomalous basins in the wave field by the common depth point method: (a) Pliocene basin (Novobogatinsk site), (b) Paleogene basin (Aksarai site).

(1) Reflectors: (A) at the base of Pliocene deposits, (I) at the base of Paleogene deposits, (VI) salt roof, (B_1) base of salt deposits; (2) the age of the sequence; (3) faults; (4) steep slope of the salt dome.

compensate them, we observe favorable conditions here for the formation of the most complete stratigraphic sections. However, each compensated basin evolves during a rather narrow stratigraphic interval, and therefore, it is doubtful whether it is typical of the given region. Therefore, it comes as no surprise that all Mesozoic compensated basins known thus far were described in literature as "anomalous" or "superthick" (Berezutskaya, 1985; *Sverkhmoshchnyi...*, 1970; Lipatova *et al.*, 1992) and, accordingly, they are not included even in the correlation part of stratigraphic schemes, with the exception of the Paleogene sections of the Tukbai and Tyubekuduk basins and Triassic sections of the Kandyk (Khobdin) basins.

In view of the above considerations, we decided to analyze numerous accumulated geologic–geophysical materials to clarify the problem of "anomalous" sections in compensated basins. A very high thickness of certain stratigraphic subdivisions is clearly observed in time sections of compensated basins. Reflection patterns of several basin types in the wave field, recorded

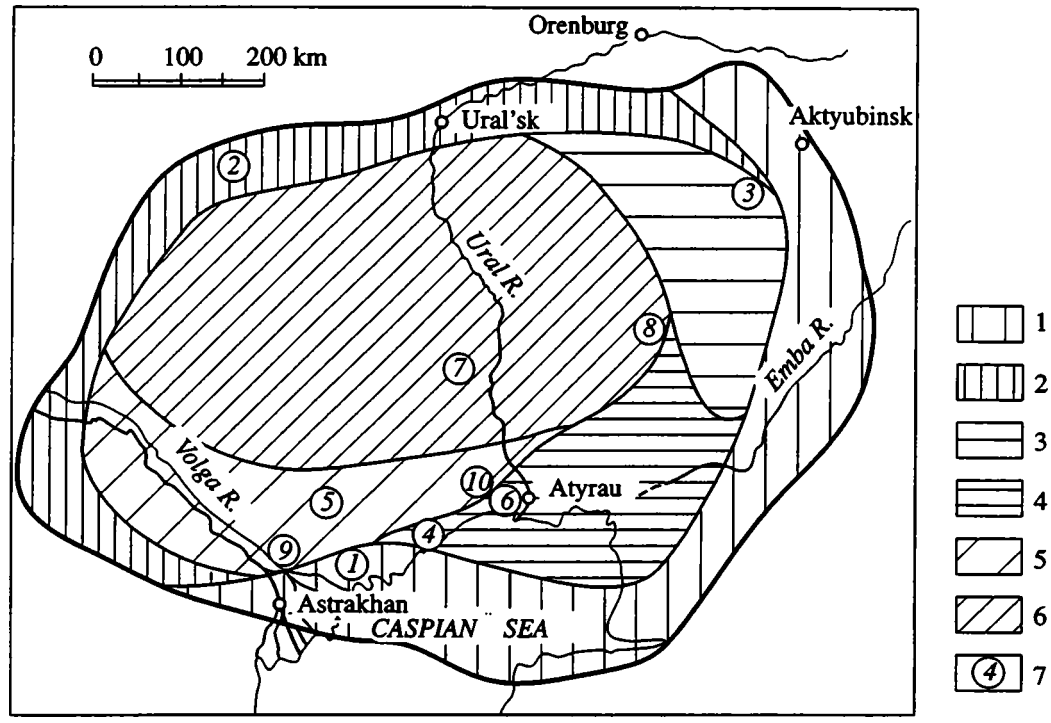

Fig. 2. Scheme of the Pericaspian depression zoning during the formation of anomalous compensated basins.
Zones with dominating age of anomalous basins:
1, Late Jurassic; 2, Late Permian–Early Triassic; 3, Triassic; 4, Triassic–Cretaceous; 5, Jurassic–Neogene; 6, Paleogene–Neogene; 7, anomalous basins. Figures in circles indicate: (*1*) Kobyakov, (*2*) South Ershov, (*3*) Khobdin, (*4*) western Sazankurak, (*5*) northern Kum, (*6*) western Kusanbai, (*7*) Tukbai, (*8*) Sorkul, (*9*) Azau, and (*10*) Novobogatinsk basins.

by the common depth point method, is given in Fig. 1. Morphologically, basins may form both local areas with increased thicknesses in the same pericline (asymmetric or one-wall basins, Fig. 1a), and isolated lenticular bodies (symmetric basins, Fig. 1b). As can be seen in the time sections, the thickness of sediments becomes larger because of the progressively increasing thickness of isolated strata and appearance of new wedged horizons. Evidently, the "basin" section was least affected by regional and local hiatuses, which commonly truncate a great part of the "dome" sections. This indicates that the compensated basin preserves the best sedimentary sequence (stratigraphically complete section) corresponding to any interval of the geological time.

The interpretation of seismic prospecting materials (seismostratigraphic analysis) enabled us to identify the diversity of compensated basins, and to figure out the time of their formation and spatial distribution in geological sections. The published materials (Berezutskaya, 1985; Koltypin, 1966; *Reshenie...*, 1982; *Sverkhmoshchnyi...*, 1970), along with the research carried out by the authors, indicate the presence of compensated basins of different age in the Peri-Caspian depression, and this makes it possible to compile nearly the entire suprasalt section.

The conducted investigations enabled us to outline a geographical zoning of the compensated basins according to the age of the deposits filling them (Fig. 2). The principal feature of zoning in the Peri-Caspian region, established by that criterion, is the concentric position of basins different in age, which suggests their sequential rejuvenation from borders towards the center of the region (in the central zone, halokinetic processes continue at present). Such zoning most likely demonstrates an increase of the primary thickness of the salt sequence in one direction and reflects the areal distribution of the main halokinesis phases in the region. The real picture is more complicated, and one zone may comprise basins different in age. In particular, A.A. Akkulov *et al.* (1992) reported, that Late Permian to Pliocene basins were discovered in the Volga–Ural southern interfluve area.

The general section of the oversalt sequence, compiled from separate sections of compensated basins, is more than three times thicker than the natural sections of interdome zones, and individual stratigraphic subdivisions differ in thickness by 5–20 times from one another. Therefore, it is important to clarify how this affects the lithofacies distribution, and whether the complete section reflects a change in the sedimentary facies, and thus allows us to regard them as "anomalous." With this purpose, we shall review several compensation basins on which drilling data are available (Fig. 3).

The Novobogatinsk basin (Column 10) is 16 × 11 km in dimension; it is limited by the Novobogatinsk,

Sugur, and Tegen salt domes, and by connecting necks. Seismic prospecting data suggest the existence of a thick (up to 3000 m) sequence of upper Pliocene-Quaternary deposits within the basin. Analogous deposits are recorded in the basin marginal parts by boreholes drilled in the 1950s. The age of rocks resting on the slopes of the salt domes was dated by organic remains as the Akchaghylian and Apcheronian (early and late). More ancient deposits were recovered above the salt domes. Along lateral faults, they contact with upper Pliocene rocks (Koltypin, 1966).

A sequence about 275 m thick, occurring between upper Pliocene and Paleogene deposits, was identified as *the Azausk basin* (Column 9) in seismic records. Characteristic top and basal unconformities suggest its early Pliocene age.

The Tukbai basin (Column 7) with dimensions about 10 × 10 km is situated in the southern Ural–Volga interfluve area, and was recovered by Hole P-24 penetrated down to a depth of 4225 m. The hole bottom is found to be in Upper Cretaceous rocks. The hole revealed a unique Paleogene–Miocene sequence: Paleocene (600 m), lower Eocene (600 m), middle Eocene (450 m), upper Eocene (196 m), Oligocene (795 m) and Miocene (1017 m). The upper Pliocene and Quaternary deposits have a thickness common for them in the Pericaspian region. The Cenozoic section, penetrated by the hole, appeared to be the thickest and stratigraphically most complete; today it is accepted as the stratotype for the Paleogene of the southern Peri-Caspian depression (*Sverkhmoshchnyi...*, 1970).

The western Kusanbai basin (Column 6) comprises an anomalous section in the Berriasian–Valanginian interval. Deep Hole P-58 in western Kusanbai penetrated 1760 m of the Berriasian and 800 m of the Valanginian deposits (the Berriasian section is 30 times thicker than a normal "dome" section recovered by Hole SP-5). The lithological composition of the rocks above the dome and in the basin is nearly the same. At the Hauterivian level, the basin's anomalous development nearly ceased, while it resumed for a short time in the Barremian. Berriasian deposits in the Peri-Caspian region were first identified in the section of Hole P-58 (Berezutskaya, 1985).

The Northern Kuma basin (Column 5) has a thick salt bench, which gave rise to the formation of a secondary salt dome. The basin was penetrated by several boreholes (G-2, G-3, G-4, and G-7). The deepest of them (G-7) penetrated 1285 m of the salt bed, the 655-m-thick carbonate Upper Jurassic sequence, the 766-m-thick Lower-Middle Jurassic deposits, and entered into the Triassic rocks. The age of the salt bench was dated as the Late Jurassic, because the rocks overlying and underlying the bench sequences contain Volgian ostracodes. Actually, the whole Jurassic section in deepest parts of the basin is anomalous as compared to its standard; even apart from the salt sequence, the Upper Jurassic anomalous sections are 1.5–2 times thicker than the normal ones.

The Sazanokurak basin (Column 4) is unique; it has the greatest thickness and most complete Upper Triassic stratigraphic section. Here, Hole G-3 recovered a thick (2471 m in the depth interval of 1600–4071 m) section of this age, properly characterized by fauna. V.V. Lipatova (All-Russia Research Geological Petroleum Institute) distinguished from bottom to top the sandy–clay deposits of the Akmalyk (900 m), Khobin (910 m), and Kusankuduk (690 m) formations. In seismic records, these units are separated by distinct reflectors, and this makes it possible to identify seismic horizons here: V_1^1, V_1^2, and V_1^3.

The lithological features of suprasalt deposits, recovered by holes, show that there is actually no lithofacial difference between the rocks filling the basins and the adjacent structures. This is especially evident from data on the Paleogene sequence. In the Pericaspian region, this complex is preserved below the pre-Pliocene washout in isolated interdome troughs, and only in the southeast of the depression does it form the sedimentary cover. However, irrespective of the thickness and stratigraphic position (Fig. 3b, Columns 7 and 8), the lithologic–facial characteristics of various Paleogene stratigraphic units are persistent. An analogous picture is observed in other stratigraphic units of the suprasalt sequence. Therefore, rock assemblages in compensated basins reflect the regional paleogeographic setting of the entire sedimentation basin, rather than local peculiarities in its particular areas. The thickness of carbonate and terrigenous rocks progressively increases in the basin, without facial changes (the Kimmeridgian beds in the northern Kuma basin, Fig. 3*C*, Column 5).

Thickness distribution patterns are complicated. The anomalously great thickness of some stratigraphic units in compensated basins is caused by halokinetic processes, which many researchers believe may significantly change depending upon local conditions. Therefore, we made a special analysis in order to clarify to what extent the rate of halokinetic movements and tectonic subsidence can be correlated. Tectonic subsidence was measured by the sedimentation rate in interdome zones, and halokinetic movements were identified by that in compensated basins. Figure 4 shows that the sedimentation rate during the Late Permian–Cenozoic notably changed at the Triassic–Jurassic and early–late Pliocene boundaries. Such a change is recorded both in standard and anomalous sections. On the whole, the curves, indicating the standard and anomalous sedimentation rates, are in agreement (the rate increase in a standard section conforms to the maximum at the anomalous value curve). The anomaly coefficient (the ratio of anomalous to standard rate values) may reach 20; for the Mesozoic–Cenozoic, it amounts on average to 3.

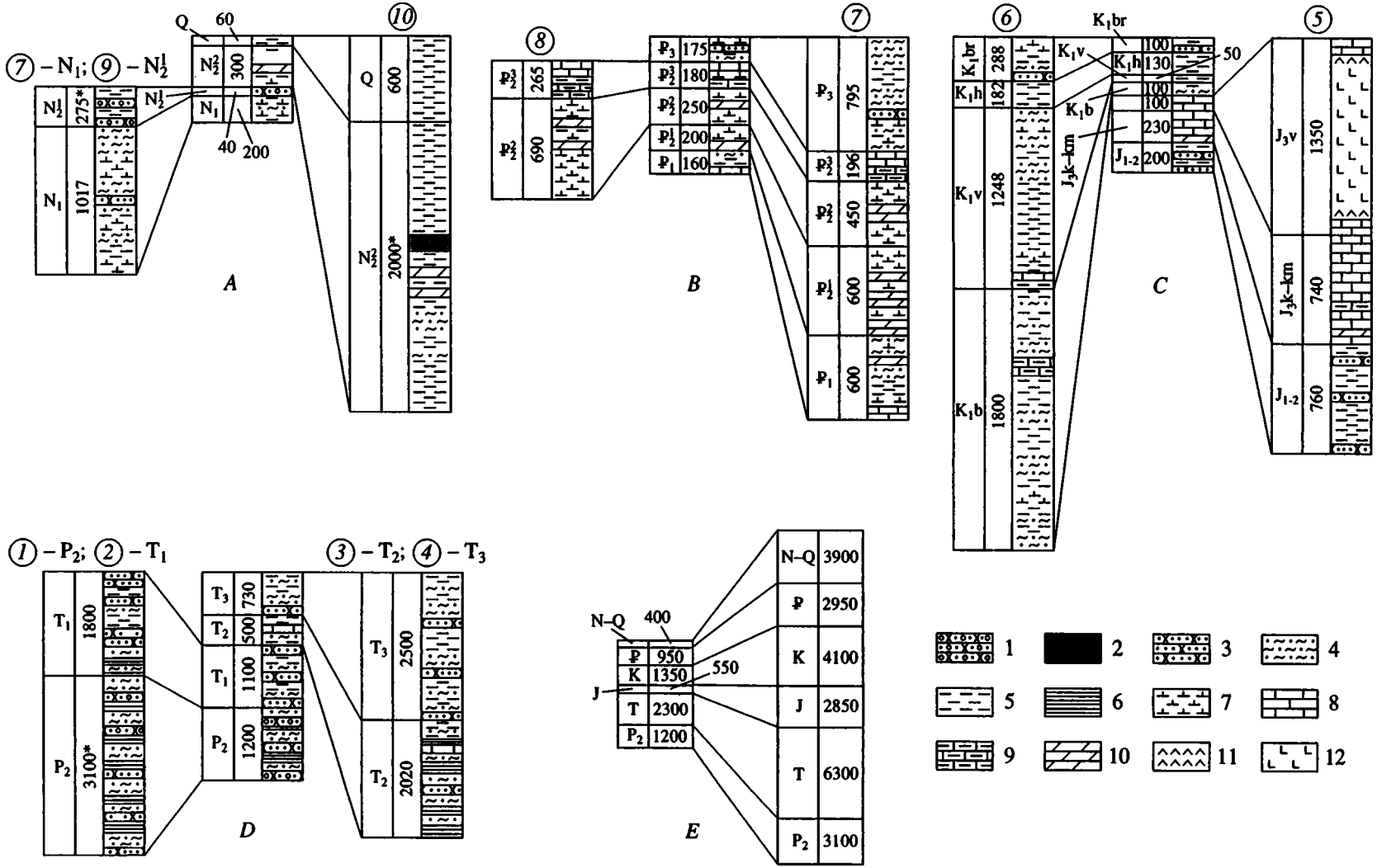

Fig. 3. Comparison of normal and anomalous basin sections in the Pericaspian region. (*A*) Neogene–Quaternary deposits, (*B*) Paleogene deposits, (*C*) Jurassic–Cretaceous deposits, (*D*) Upper Permian–Triassic deposits (normal sections of *A*–*D* are shown in the center), (*E*) composite normal (left) and anomalous sections. Figures in circles correspond to geographic localities in Fig. 2.

Lithology: 1, gravelstone, conglomerate; 2, sand; 3, sandstone; 4, siltstone; 5, clay; 6, claystone; 7, calcareous clay; 8, limestones; 9, argillaceous limestone; 10, marl; 11, anhydrite; 12, halite.

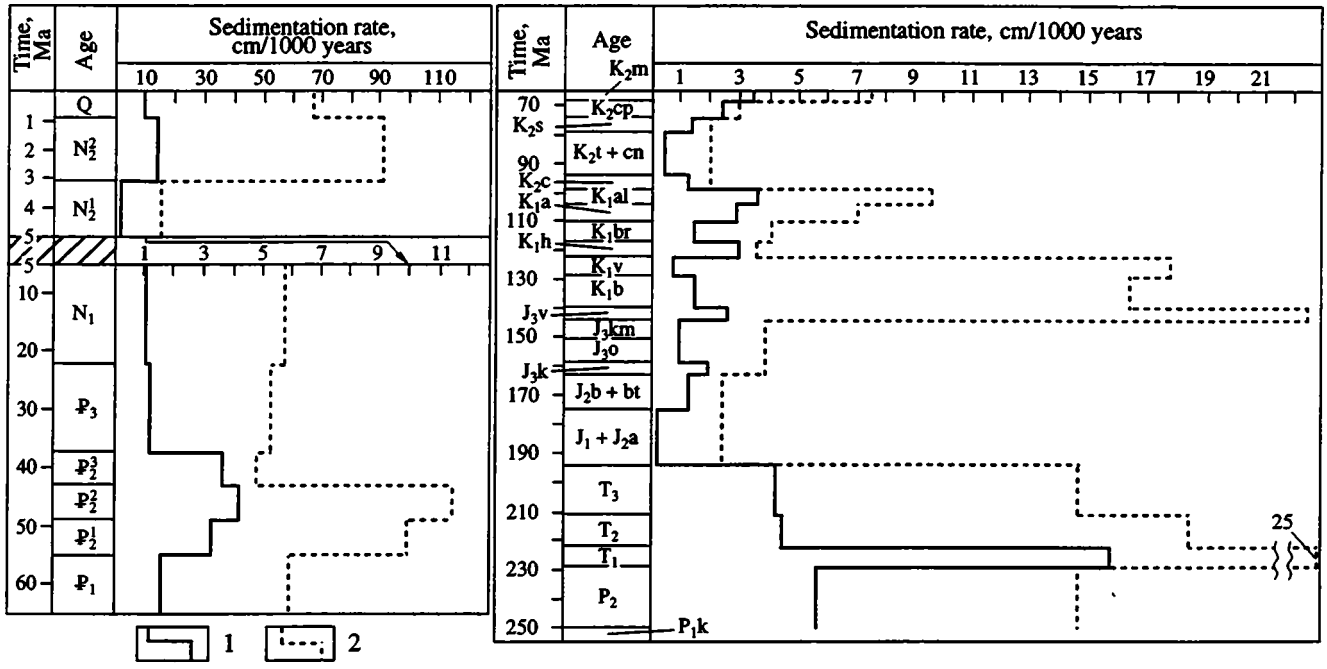

Fig. 4. Diagrams of sedimentation rates in suprasalt deposits of the Peri-Caspian depression: (1) normal sedimentation rates curve; (2) anomalous sedimentation rates curve.

Provided that the average rate of the Mesozoic–Cenozoic sedimentation in a standard section is 1–3 cm/1000 years, several peaks are distinguished: Lower Triassic (15.5 cm), Maestrichtian (7.5 cm), middle Eocene (4.5 cm), Pliocene (14 cm), and Quaternary (10 cm).

Geodetic investigations at the Elton and Baskunchak domes suggest the rate of their growth to be about 1 mm/year, or 100 cm/1000 years (Svitoch, 1987). This value agrees well with sedimentation rates in the Pliocene–Quaternary anomalous basins (70–90 cm/1000 years). If the dome lifting rate (and correspondingly, subsidence of the compensation basin) is 1 mm/year, one million years is required for the basin to fill up with 1-km-thick sediments; 2-km-thick sediments will fill up the basin in 2 m.y., and it will take 3 m.y. to fill up the basin with 3-km-thick sediments, etc. (the maximum rate of halokinetic subsidence). Commonly, the studied stratigraphic range embraces a much broader interval of geological time.

Actually, the sedimentation rate in anomalous basins is not always close to maximum. This is due to: (1) regional and local unconformities, which eliminate some deposits that would otherwise be considered in calculations; (2) the insufficient thickness of primary salt, which prevented the subsidence of a compansated basin to the depth corresponding to its age range; and (3) the presence of sediments deposited earlier in primary basins that restrained the subsidence of secondary basins to a substantial depth. There may exist a mechanism, which we called "the basin anomaly limitation," that prevents the long-term formation of a basin with a constant depositional center. On average, the life of anomalous basins is 15–20 m.y. Among the exceptions, there is the short-term Novobogatinsk basin (about 5 m.y.) and the long-lived Tukbai basin (more than 50 m.y.). Figure 5 also indicates that anomalies have quite clear boundaries; occasionally, thick sections are separated from normal sequences by stratigraphic intervals with sections of a transitional type.

Comparing the sedimentation rate curves for anomalous and normal sections, we can conclude that despite considerable variations in absolute values, both curves are similar. Consequently, halokinetic movements, which control the formation of compensated basins, do not alter the general sedimentary process in time to any considerable degree.

CONCLUSION

Thus, the conclusions of the research are as follows:

(1) Salt tectogenesis in the Peri-Caspian depression gave rise to the formation of unique structures: secondary (compensated) basins with an anomalous thickness of the stratigraphic subdivisions of a narrow stratigraphic range.

(2) Compensated basins are distributed as generally concentric zones, where filling sequences are progressively younger from borders (Upper Permian basins) towards the center (Quaternary) of the original depression.

(3) Comparison of seismic and log data suggests that the increase of sedimentary thickness is a result of swelling in certain strata and the appearance of new beds absent in the "dome" section.

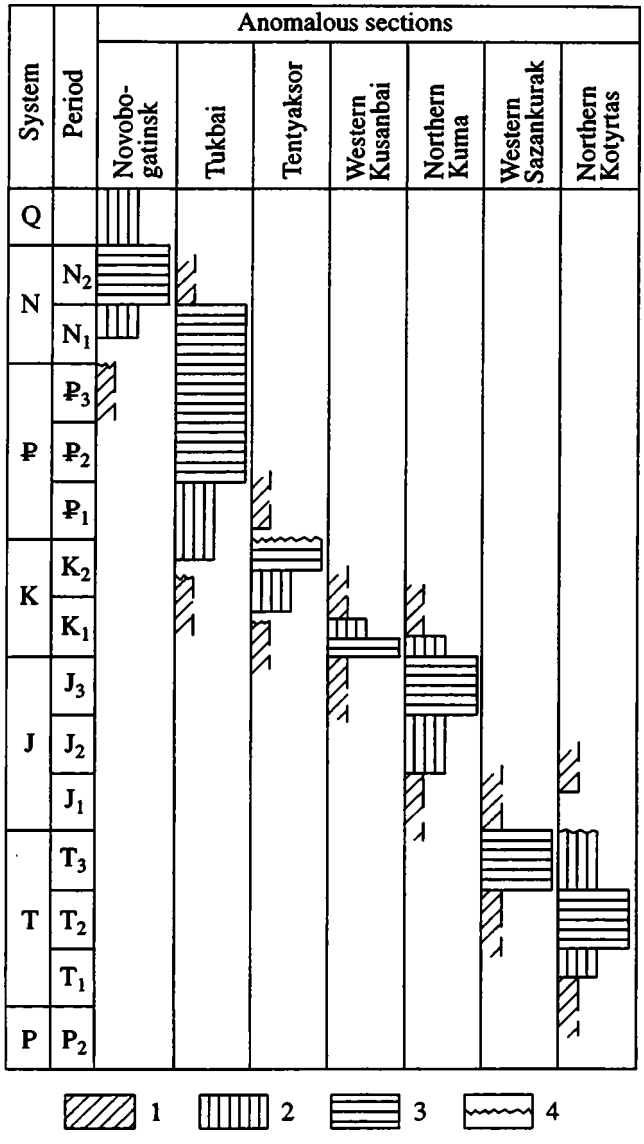

Fig. 5. Duration of compensated basin development: (1) standard thickness; (2) elevated thickness; (3) anomalous thickness; (4) washout.

(4) Interpretation of log curves and lithological characteristics of the basin and neighbouring dome sections suggest that the lithologic-facial composition of coeval deposits does not actually change, and mostly depends on the regional paleogeographical environment.

(5) Sedimentation rates in anomalous basins are 3–20 times higher than in normal structures, though, on the whole, the curves of anomalous and normal sedimentation rates are similar.

(6) The analysis shows that sedimentary sections formed in compensated basins should be used as stratotype sections of local stratigraphic subdivisions.

On the whole, the full age spectrum of compensated basins (from Late Permian to Pliocene inclusive) may be used for the detalization of both the local lithostratigraphic and regional biostratigraphic schemes. A special study on these sections will make it possible to discover the missing stratigraphic subdivisions. In connection, it would be reasonable to organize an international geostratigraphic site in the Pericaspian depression in order to study the entire record of suprasalt deposits.

The exploration of compensated basins will greatly contribute to industrial oil and gas activities, as the thickness of "classic" productive horizons might be markedly enlarged, and wide gradients of thickness provide favorable conditions for the genesis of nonanticlinal (lithological) traps.

Reviewed by Yu.B. Gladenkov

ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research, project no. 14768.

REFERENCES

Akkulov, A.A., Turkov, O.S., and Shudabaev, K.S., New Data on the Structure and Formation Time of Interdome Zones in the South of the Peri-Caspian Depression, *Geol. Nefti Gaza*, 1992, no. 4, pp. 2–4.

Berezutskaya, L.V., Extremely Thick Lower Neocomian Section in the Peri-Caspian Region (Western Kusanbai), *Byull. Mosk. O-va Ispyt. Prir., Otd. Geol.*, 1985, vol. 60, issue 4, pp. 68–73.

Koltypin, S.N., Deep-Seated Interdome Basins in the South of the Peri-Caspian Depression, in *Usloviya obrazovaniya i osobennosti neftegazonosnosti solyanokupol'nykh struktur* (Formation Conditions and Specific Features of Oil–Gas Occurrences in Salt Dome Structures), Kiev: Naukova Dumka, 1966, vol. 1, pp. 127–136.

Lipatova, V.V., Pikalova, O.V., Basilova, S.K., *et al.*, Superthick Section of the Masteksai Horizon (Middle Triassic) in the Southern Ural-Volga Interfluve Area, *Geol. Nefti Gaza*, 1992, no. 9, pp. 19–22.

Reshenie mezhvedomstvennogo stratigraficheskogo soveshchaniya po triasu Vostochno-Evropeiskoi Platformy (The Resolution of Interdepartmental Meeting on the Triassic of the East European Platform), Leningrad: Vses. Geol. Inst. (VSEGEI), 1982.

Svitoch, A.A., An Underground Phenomenon of the Peri-Caspian Depression, *Priroda*, 1987, no. 7, pp. 61–65.

Stratigraficheskii kodeks SSSR (The Stratigraphic Code of the USSR), Zhamoida, A.I., Kovalevskii, O.P., Moiseeva, A.I., and Yarkin, V.I., Eds., Leningrad: Vses. Geol. Inst. (VSEGEI), 1977.

Vasil'ev, Yu.M., Gibshman, N.B., Li, K.A., *et al.*, Superthick Cenozoic Section in the Center of the Peri-Caspian Salt Dome Area, in *Poiski nefti i gaza v solyanokupol'nykh oblastyakh* (Prospecting for Oil and Gas in Salt Dome Areas), Moscow: Nedra, 1970, issue 90, pp. 135–142.

Zhuravlev, V.S., *Sravnitel'naya tektonika Pechorskoi, Prikaspiiskoi i Severomorskoi ekzogonal'nykh vpadin Evropeiskoi platformy* (Comparative Tectonics of the Pechora, Peri-Caspian, and Severomorsk Exogonal Depressions in the European Platform), Moscow: Nauka, 1972.

Stratigraphy and Geological Correlation, Vol. 4, No. 4, 1996, pp. 416–418. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 4, No. 4, 1996, pp. 109–111.
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A Chapter in the History of Paleontology and Stratigraphy in Russia

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Received February 9, 1995

The first large class of Russian geologists graduated from Leningrad University in 1929 (admitted in 1923). Twelve graduates among them, including O.S. Vyalov, G.Ya. Krymgol'ts, N.P. Luppov, V.S. Slodkevich, and A.G. Eberzin, specialized in stratigraphy and paleontology with remarkable success.

Our fellow graduates formed a group of student geologists, who were admitted in 1923 into the Natural History Department of the Physical–Mathematical Faculty of Petrograd University; later, several students from other higher educational establishments joined our batch, which brought its number to 30. At the end of the nineteenth and beginning of the twentieth centuries, only a few of the Natural History graduates from this faculty specialized in geology, and our group was known as the first mass admission into, and subsequently, the first mass graduation of geologists from the university.

In the early 1920s, the demand for geologists was keenly felt, because the mining colleges that turn out mining engineers were unable to satisfy it. This challenge accounts for the interest of students in geological sciences, in which the geological school of the Leningrad State University has an impressive record.

In the 1920s, the position of geology in the structure of the university changed for the better, as did the curriculum and training of future geologists. By the time of our graduation from the Physical–Mathematical Faculty, the Natural History Department expanded to form the Geological–Mineralogical Department, with geological, mineralogical, paleontological, and other subdivisions (further referred to as the Geological Department). The lecturers were world-famous scientists, such as Ya.S. Edel'shtein (general geology), M.E. Yanishevskii (paleontology), S.M. Kurbatov (mineralogy), P.A. Pravoslavlev (historical geology), F.Yu. Levinson-Lessing (petrography and ore deposits), and A.N. Krishtofovich (paleobotany). At the head of the Department were Professor P.A. Pravoslavlev and Associate Professor V.F. Pchelintsev.

The university regulations of that time allowed free attendance of lectures by students, which is still the policy to date. In order to complete each year of study, the students were not obliged to pass examinations on all the disciplines of the previous year; rather, they could take examinations during the entire academic course. For example, G.Ya. Krymgol'ts passed his last physics exam from his second year just before presenting his graduation thesis. Moreover, in 1929, the graduation thesis requirement was canceled for a period. Only three of us, O.S. Vyalov, N.P. Luppov, and A.G. Eberzin, presented theses and were the first to graduate from the university in 1928. These liberal policies were particularly welcome, as they allowed all students to concurrently work in geological institutions, an experience which often determined their future career in geology.

It was actually during this time that advances in stratigraphy and paleontology were launched in Russia; later, many of our fellow graduates worked in stratigraphy and related paleontology, the two fundamental geological sciences—twelve out of the twenty students in the photograph taken in 1929. It is no exaggeration to state that they appreciably contributed to our knowledge on the numerous fossil faunas and to the development of stratigraphy, particularly that of the Mesozoic and Cenozoic, in the different regions over the vast territory of Russia.

Mentioned below are our fellow graduates of the first mass graduation, who have dedicated their work to stratigraphy and paleontology.

Boris Nikolaevich Aver'yanov (1906–1941/42 ?). One of the first Russian specialists on graptolites and author of works on hemichordates from Ordovician and Silurian deposits of the Urals, Novaya Zemlya, Central Asia, and northeastern Russia.

Oleg Stepanovich Vyalov (1904–1988). Academician of the Academy of Sciences of Ukraine, a specialist in many branches of geological sciences, an indomitable traveller, and the first Russian geologist to work in Antarctica. Author of more than 500 books and papers. In stratigraphy and paleontology, he is best known as the originator of the Paleogene stratigraphic scale of Central Asia and of the ostreids system, and as the founder of ichnology, a new trend in paleontology.

† Deceased.

[1] T.N. Bogdanova and S.V. Lobachova (All-Russia Research Institute of Geology, VSEGEI), S.O. Vyalova (Library, Russian Academy of Sciences, BAN, St. Petersburg), and V.A. Prozorovskii (St. Petersburg State University) have rendered valuable assistance in the preparation of this paper.

Sitting from left to right: Prokof'ev, G.T. Petrova (Pchelintseva), V.S. Slodkevich, O.S. Vyalov, E.I. Sokolova, A. Smirnova.

Standing, lower row: L.V. Romanovskaya, A.P. Il'ina, T.A. Mordvilko, A.G. Eberzin, N.N. Lobashova, O.S. Gracheva, G.Ya. Krymgol'ts.

Standing, upper row: N.V. Avchinnikov, K.P. Pirogov, N.P. Luppov, G.G. Lemmlein, V.S. Ser'myagin, I.F. Pustovalov, B.N. Aver'yanov.

Agniya Petrovna Il'ina (born 1904). A specialist in Cenozoic bivalves and nummulites, and in the Paleogene and Neogene stratigraphy of Mangyshlak, Crimea, and the Far East.

Grigorii Yakovlevich Krymgol'ts (born 1907). Professor of Leningrad (now St. Petersburg) University, a renown specialist in Jurassic stratigraphy and in cephalopod mollusks. Chair of the Commission on Jurassic Stratigraphy of the Interdepartmental Stratigraphic Committee (ISC) from its institution until 1987.

Nikolai Pavlovich Luppov (1904–1975). Professor, Honored Scientist of Turkmenistan, a distinguished expert on Central Asian geology and stratigraphy of Cretaceous deposits, Cretaceous ammonites and bivalves. Scientific Supervisor (1956–1966) of the VSEGEI Expedition in Central Asia. From 1964 until his death, Chair of the ISC Commission on the Stratigraphy of the Cretaceous System.

Tat'yana Aleksandrovna Mordvilko (1905–1985). An outstanding specialist on the Early Cretaceous bivalve mollusks, author of the unique Catalog of the Lower Cretaceous Pelecypods of the Southern USSR (unfortunately, only a part of the Catalog was published). A distinguished specialist on the lower Cretaceous stratigraphy of the Caucasus and Central Asia.

Galina Timofeevna Petrova (Pchelintseva) (1906–1958). A well-known specialist on Jurassic bivalves from the southern and eastern periphery of Asia.

Lidiya Vasil'evna Romanovskaya (1902–1989). Head of the Department of Paleontological Collections of TsNIGRI (Central Scientific Research Institute for Geological Prospecting) at the F.N. Chernyshev Museum. Before the Second World War, Romanovskaya successfully studied the Late Cretaceous bivalve mollusks of Sakhalin, Central Asia, and Donbass.

Vsevolod Sergeevich Slodkevich (1904–1964). Professor, Honored Scientist of the Karelian–Finnish Republic, an outstanding organizer of scientific research in Kareliya, Siberia, and the Far East, a renown expert on Paleogene stratigraphy and mollusks.

Anastasiya Smirnova (1903–1932). Studied the Cenozoic bivalves of Sakhalin.

Ekaterina Ivanovna Sokolova (1903–1990). A specialist on the Permian and Mesozoic stratigraphy and the geology of oil-bearnig deposits in the Peri-Caspian syneclise.

Anatolii Georgievich Eberzin (1904–1970). Professor, Head of Laboratory of the lowest mollusks at the Paleontological Institute, Academy of Sciences of the USSR; one of the leading specialists in mollusks, stratigraphy, and paleogeography of the Neogene. Head of the ISC Commission on Stratigraphy of the Neogene System from its foundation.

Our fellow graduates named above greatly contributed to the paleontological–stratigraphical science in Russia with their numerous original monographs and papers, most of which are still scientifically relevant. The works of O.S. Vyalov, G.Ya. Krymgol'ts, N.P. Luppov, V.S. Slodkevich, A.G. Eberzin, and others constitute parts of the volumes of *Stratigraphiya SSSR* (Stratigraphy of the USSR) devoted to the Jurassic, Cretaceous, Paleogene and Neogene systems; their papers were included in the volumes of *Paleontologiya SSSR* (Paleontology of the USSR) (cephalopods, gastropods and bivalve mollusks), *Paleogeograficheskie Atlasy* (Paleogeographic Atlases) and other summarizing publications, of which they were the authors and editors.

The photograph of the first graduates also shows I.F. Pustovalov and V.S. Ser'myagin, geologists and lithologists; O.S.Gracheva, G.G. Lemmlein, N.N. Lobashova, and Prokof'ev, geochemists; N.V. Avchinnikov and K.P. Pirogov, hydrogeologists.

We should mention that Oleg Stepanovich Vyalov was the initiator of the idea to publish this group portrait and comments, though he did not live to see its realization.

Reviewer B.S. Sokolov

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